



# Measurement report: Nitrogen isotopes ( $\delta^{15}\text{N}$ ) and first quantification of oxygen isotope anomalies ( $\Delta^{17}\text{O}$ , $\delta^{18}\text{O}$ ) in atmospheric nitrogen dioxide

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**Abstract.** The isotopic composition of nitrogen and oxygen in nitrogen dioxide ( $\text{NO}_2$ ) potentially carries a wealth of information about the dynamics of the nitrogen oxides ( $\text{NO}_x = \text{nitric oxide (NO)} + \text{NO}_2$ ) chemistry in the atmosphere. While nitrogen isotopes of  $\text{NO}_2$  are subtle indicators of  $\text{NO}_x$  emissions and chemistry, oxygen isotopes are believed to reflect only the  $\text{O}_3 / \text{NO}_x / \text{VOC}$  chemical regime in different atmospheric environments. In order to access this potential tracer of the tropospheric chemistry, we have developed an efficient active method to trap atmospheric  $\text{NO}_2$  on denuder tubes and measured, for the first time, its multi-isotopic composition ( $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$ ). The  $\Delta^{17}\text{O}$  values of  $\text{NO}_2$  trapped at our site in Grenoble, France, show a large diurnal cycle peaking in late morning at  $(39.2 \pm 0.3)\text{‰}$  and decreasing at night until  $(20.5 \pm 0.3)\text{‰}$ . On top of this diurnal cycle,  $\Delta^{17}\text{O}$  also exhibits substantial daytime variability (from  $29.7\text{‰}$  to  $39.2\text{‰}$ ), certainly driven by changes in the  $\text{O}_3$  to peroxy radicals ( $\text{RO}_2$ ) ratio. The nighttime decay of  $\Delta^{17}\text{O}(\text{NO}_2)$  appears to be driven by  $\text{NO}_2$  slow removal, mostly from conversion into  $\text{N}_2\text{O}_5$ , and its formation from the reaction between  $\text{O}_3$  and freshly emitted  $\text{NO}$ . As expected from a nighttime  $\Delta^{17}\text{O}(\text{NO}_2)$  expression, our  $\Delta^{17}\text{O}(\text{NO}_2)$  measured towards the end of the night is quantitatively consistent with typical values of  $\Delta^{17}\text{O}(\text{O}_3)$ . Daytime N isotope fractionation is estimated using a general expression linking it to  $\Delta^{17}\text{O}(\text{NO}_2)$ . An expression is also derived for the nighttime N isotope fractionation. In contrast to  $\Delta^{17}\text{O}(\text{NO}_2)$ ,  $\delta^{15}\text{N}(\text{NO}_2)$  measurements exhibit little diurnal variability ( $-11.8\text{‰}$  to  $-4.9\text{‰}$ ) with negligible isotope fractionations between  $\text{NO}$  and  $\text{NO}_2$ , mainly due to

high  $\text{NO}_2 / \text{NO}_x$  ratios, excepted during the morning rush hours. The main  $\text{NO}_x$  emission sources are estimated using a Bayesian isotope mixing model, indicating the predominance of traffic emissions in this area. These preliminary results are very promising for using the combination of  $\Delta^{17}\text{O}$  and  $\delta^{15}\text{N}$  of  $\text{NO}_2$  as a probe of the  $\text{NO}_x$  sources and fate and for interpreting nitrate isotopic composition records.

## 1 Introduction

Nitrogen oxides ( $\text{NO}_x = \text{NO}_2 + \text{NO}$ ) are at the heart of tropospheric chemistry, as they are involved in key reaction chains governing the production and destruction of compounds of fundamental interest for health, ecosystems, and climate issues (Brown, 2006; Finlayson-Pitts and Pitts, 2000; Jacob, 1999). For example,  $\text{NO}_2$  photolysis followed by reaction of  $\text{NO}$  with peroxy radicals ( $\text{RO}_2 = \text{HO}_2 + \text{RO}_2$ ) is the only significant source of ozone ( $\text{O}_3$ ) in the troposphere, where it serves as a severe air pollutant and a greenhouse gas. Tropospheric  $\text{O}_3$  also plays a major role in the production processes of radicals which are responsible for the oxidation and removal of compounds emitted into the atmosphere (Crutzen, 1996). This “cleaning” ability is referred to as the atmospheric oxidative capacity (AOC; Prinn, 2003). Additionally,  $\text{NO}_x$  species are at the core of the reactive nitrogen cycle as precursors of atmospheric nitrate (particulate  $\text{NO}_3^-$  + gaseous  $\text{HNO}_3$ ), which contributes to soil acidification and eutrophication (Galloway et al., 2004), as well as aerosol radiative forcing (Liao and Seinfeld, 2005). In order

to better understand the reactive nitrogen (which includes  $\text{NO}_x$  and  $\text{HNO}_3$ ) chemistry, the related AOC, and the contributions of precursor emissions to nitrate deposition, it is necessary to better constrain  $\text{NO}_x$  emission sources and individual oxidation processes.

Stable isotope analysis is a powerful tool for tracing emission sources, individual chemical mechanisms, and budgets of atmospheric trace gases (Kaye, 1987). Because physico-chemical and biological processes favour lighter or heavier isotopologues, the isotopic composition of a chemical species will often vary according to its formation pathway. This phenomenon of isotopic fractionation can thus be used to trace different processes involved in the formation of the chemical species being analysed. Isotopic enrichment ( $\delta$ ) of an element  $X$  is expressed in per mill (‰) and defined as  $\delta^X = ({}^nR_{\text{spl}}/{}^nR_{\text{ref}} - 1)$  with  ${}^nR$  being the elemental isotope abundance ratio of the heavy isotope over the light isotope (e.g. for oxygen isotopes  ${}^{18}R({}^{18}\text{O}/{}^{16}\text{O}) \equiv {}^{18}R = x({}^{18}\text{O})/x({}^{16}\text{O})$  or  ${}^{17}R({}^{17}\text{O}/{}^{16}\text{O}) \equiv {}^{17}R = x({}^{17}\text{O})/x({}^{16}\text{O})$ , with  $x$  being the isotopic abundance) in a sample ( ${}^nR_{\text{spl}}$ ) and in a reference ( ${}^nR_{\text{ref}}$ ). The Vienna Standard Mean Ocean Water (VSMOW; Li et al., 1988) and atmospheric nitrogen ( $\text{N}_2$ ; Mariotti, 1984) are the international references for oxygen and nitrogen ratios, respectively. Most natural isotopic fractionations are mass-dependent fractionations (MDFs; Urey, 1947), as is notably the case for terrestrial oxygenated species in which the triple oxygen composition follows  $\delta^{17}\text{O} \approx 0.52 \times \delta^{18}\text{O}$  (Thiemens, 1999). Yet, laboratory experiments (Thiemens and Heidenreich, 1983) and atmospheric observations (Johnston and Thiemens, 1997; Krankowsky et al., 1995; Vicars and Savarino, 2014) have shown that the isotopic composition of ozone formed in the atmosphere does not follow this canonical MDF relationship and reflects mass-independent fractionation (MIF) processes. The important deviation from the MDF oxygen relationship is called the oxygen-17 anomaly ( $\Delta^{17}\text{O}$ ) and is defined here in its approximate linearized form as  $\Delta^{17}\text{O} = \delta^{17}\text{O} - 0.52 \times \delta^{18}\text{O}$ . Our choice of this linear definition is mainly motivated by its convenience for mass-balance calculations and its validity for our large  $\Delta^{17}\text{O}$  values and variability. Overall, biases related to our choice of the linear definition are marginal in our conditions (Assonov and Brenninkmeijer, 2005). It follows that  $\Delta^{17}\text{O}$  inherited from ozone can be considered conserved during MDF processes.

The multi-isotopic composition of  $\text{NO}_x$  is therefore a very valuable tracer of its emissions and chemistry in the atmosphere. However, so far,  $\Delta^{17}\text{O}$  of atmospheric  $\text{NO}_2$  ( $\Delta^{17}\text{O}(\text{NO}_2)$ ) has been investigated only using laboratory (Michalski et al., 2014) and modelling (Alexander et al., 2020, 2009; Lyons, 2001; Morin et al., 2011) approaches with theoretical frameworks, and these results need to be tested against atmospheric observations. Walters et al. (2018) have presented a method of sampling and analysing nitrogen and oxygen stable isotopes of  $\text{NO}_2$  collected separately at

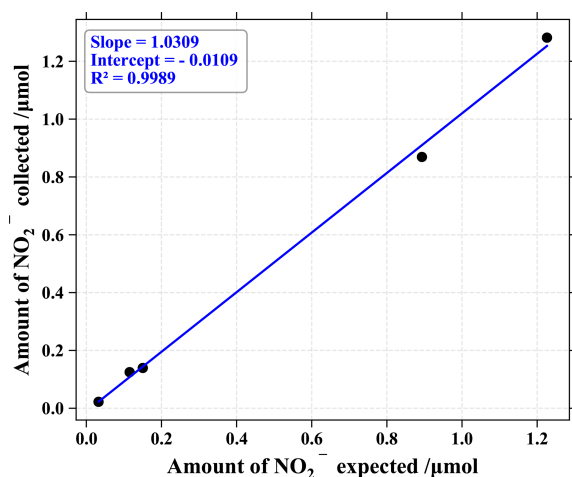
daytime and nighttime in an urban area, but they did not report on  $\Delta^{17}\text{O}$ . Dahal and Hastings (2016) have attempted to measure  $\Delta^{17}\text{O}$  of  $\text{NO}_2$  collected on passive samplers, but the isotopic signal was partly degraded during the sampling and the analytical procedure. Building on their work, we present here an efficient method to collect atmospheric  $\text{NO}_2$  for isotopic analysis and present the first measurements of triple oxygen isotopes and double nitrogen isotopes of atmospheric  $\text{NO}_2$ . Combined with mass-balance equations, oxygen isotopes are used to investigate the links between the variability of the oxygen isotope anomaly of  $\text{NO}_2$  and its formation pathways. We also revisit the Morin et al. (2011)  $\text{NO}_x$  isotopic theoretical framework and extend it to urban environments. After estimating the nitrogen isotopic fractionation between  $\text{NO}$  and  $\text{NO}_2$ , we infer from  $\delta^{15}\text{N}$  of  $\text{NO}_2$  ( $\delta^{15}\text{N}(\text{NO}_2)$ ) the major emission sources of  $\text{NO}_x$  influencing our sampling site using an isotopic mixing model (Parnell et al., 2010).

## 2 Materials and methods

### 2.1 Sampling method

$\text{NO}_2$  was sampled on an active (pumped) collection system using denuder tubes. This method is more efficient to collect  $\text{NO}_2$  than passive methods (Røyset, 1998), allowing for shorter collection times with a breakthrough of the absorption capacity below 1 % (Buttini et al., 1987; Williams and Grosjean, 1990). The sampled air was pumped through a ChemComb® 3500 speciation cartridge (Thermo Scientific®, USA). Initially used for the speciation of gases and aerosols, these advanced sampling platforms consist of a  $\text{PM}_{2.5}$  impactor inlet connected to a stainless-steel cylinder that contains two glass honeycomb denuders connected in series for gas collection and a Teflon stage filter pack for aerosols. To collect  $\text{NO}_2$ , glass tubes were coated with an alkaline guaiacol solution. In basic medium, guaiacol (IUPAC name: 2-methoxyphenol) is known to react with  $\text{NO}_2$  to form stable  $\text{NO}_2^-$  ions (Nash, 1970), preserving the original  $\text{NO}_2$  isotopic signal due to the basic nature of the medium ( $\text{pH} = 14$  after 10 mL extraction). Because  $\text{NO}$  or peroxyacetyl nitrate (PAN) are not collected by guaiacol, this methodology avoids potential interferences from these compounds in later analyses (Buttini et al., 1987). Although nitrous acid ( $\text{HONO}$ ) can bind as  $\text{NO}_2^-$ , it is unlikely to adversely impact the results as its mixing ratio (i.e. mole fraction) is much lower than  $\text{NO}_2$  (by a factor of 10 to 20) even in very polluted cities (e.g. Harris et al., 1982; Michoud et al., 2014; Huang et al., 2017).

To evaluate the sampling system performance, a series of experiments were run with artificial gaseous  $\text{NO}_2$ . Using a commercial gas-standard generator (KinTek FlexStream®) feed with zero air, diluted  $\text{NO}_2$  (Metronics Dynacal®) was sent through a ChemComb cartridge, while  $\text{NO}_x$  mixing ratio was measured upstream and downstream of the car-



**Figure 1.** Correlation plot of  $\text{NO}_2$  collected on the first denuder tube of the sampling cartridge vs.  $\text{NO}_2$  produced by the gas standard generator.

tridge. From 1 to  $30 \text{ nmol mol}^{-1}$  of  $\text{NO}_2$  (representative of rural to urban atmospheric conditions), mixing ratios coming out of the cartridge were never above the noise level of the  $\text{NO}_x$  monitor ( $2.5 \text{ nmol mol}^{-1}$ ). To estimate the denuder's trapping efficiency, we passed different mixing ratios of gaseous  $\text{NO}_2$  through the collection apparatus and measured the amount of  $\text{NO}_2^-$  collected on the two denuders both connected in series. The denuder efficiency  $E$  was then calculated according to the following equation (Buttini et al., 1987):

$$E = \left(1 - \frac{b}{a}\right) \times 100\%, \quad (1)$$

with  $a$  and  $b$  representing the amount of  $\text{NO}_2^-$  collected on the first and the second denuder, respectively. From 0.3 to  $1.3 \mu\text{mol}$  of generated  $\text{NO}_2$  (see Fig. 1), the mean  $E$  value was about  $(97 \pm 3)\%$ . The amount of  $\text{NO}_2^-$  measured on the second denuder was reproducible and equivalent to blanks, representing on average 3 % of the quantity measured on the first denuder. In light of these results, the denuder in the second position was not subjected to isotopic analysis and allowed for trapping efficiency control.

## 2.2 Isotopic analysis

Simultaneous isotopic analyses of  $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ , and  $\delta^{17}\text{O}$  were performed using a Finnigan<sup>®</sup> MAT253 isotope ratio mass spectrometer (IRMS) following techniques described by Casciotti et al. (2002) and Kaiser et al. (2007). The azide method (McIlvin and Altabet, 2005) was used with  $\approx 100 \text{ nmol}$  of nitrites converted to  $\text{N}_2\text{O}$  using a 50 : 50 by volume mixture of 2 M sodium azide and 100 % acetic acid. This chemical method has the advantage over the bacterial method to be free of nitrate interferences since  $\text{HNO}_3$  is certainly trapped with  $\text{NO}_2$  in the basic solution coating of the

denuder tube. The principle of identical treatment (Brand, 1996) was strictly respected where the standards and samples possessed the same nitrite amount, water isotopes, total volume and matrix. Three international  $\text{KNO}_2$  salt standards – RSIL-N7373, RSIL-N10219, and RSIL-N23 with respective  $\delta^{15}\text{N} / \delta^{18}\text{O}$  values of  $-79.6/4.2\text{‰}$ ,  $2.8/88.5\text{‰}$ , and  $3.7/11.4\text{‰}$  – were used for normalization of the  $\delta$  scale. Scale contraction factors were obtained with the linear regression between measured and known values of  $\delta^{15}\text{N}$  and  $\delta^{18}\text{O}$ . Although the three standards cover a wide range of isotopic composition in  $\delta^{15}\text{N}$  and  $\delta^{18}\text{O}$ , they do not have an isotopic anomaly in  $^{17}\text{O}$ . For the  $\delta^{17}\text{O}$  scale, MDF fractionation slope (0.52) is assumed for two of these laboratory-prepared nitrite standards (see Appendix A for more details). Accuracy of this analytical method on  $\delta^{17}\text{O}$ ,  $\delta^{18}\text{O}$ , and  $\delta^{15}\text{N}$  measurements was estimated as the standard deviation ( $\sigma$ ) of the residuals between our measurements of the RSIL standards and their expected values. Additionally, isotopic integrity from denuder's extraction to the analysis by IRMS has been investigated and showed no degradation over several weeks (see Appendix B), confirming that this method is suitable for isotopic analysis of  $\text{NO}_2$ , as first demonstrated by Walters et al. (2018). The uncertainties applied to our measurements of  $\delta^{15}\text{N}$ ,  $\delta^{17}\text{O}$ , and  $\delta^{18}\text{O}$  are reported as the propagation error of the measurement uncertainty and the uncertainty resulting from sample storage. Uncertainty on  $\Delta^{17}\text{O}$  is derived from the propagation error of the overall uncertainty on  $\delta^{17}\text{O}$  and  $\delta^{18}\text{O}$ . In our study, average uncertainties on  $\delta^{15}\text{N}$ ,  $\delta^{17}\text{O}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$  are estimated to be  $\pm 0.1\text{‰}$ ,  $\pm 0.8\text{‰}$ ,  $\pm 1.8\text{‰}$ , and  $\pm 0.3\text{‰}$ , respectively ( $1\sigma$  uncertainties).

## 2.3 Study site and atmospheric $\text{NO}_2$ collection

Atmospheric  $\text{NO}_2$  was collected at the Université Grenoble Alpes campus site. Located in the eastern Grenoble urban area (690 000 inhabitants), the campus stands between a major transportation route and the Isère river. The city is located at the confluence of three valleys surrounded by mountain chains that influence the atmospheric dynamics and the local air quality. During winter, persistent temperature inversions combined with intense domestic heating can lead to severe  $\text{PM}_{10}$  pollution events (Largerone and Staquet, 2016) with daily-average mixing ratio above World Health Organization thresholds. In summer, emissions are mainly controlled by road traffic that can result in heightened ozone mixing ratios, especially during stagnant conditions.

Samplings were conducted on a platform 5 m above the ground surface. Ambient air was drawn through the cartridge with a Millipore vacuum pump at a flow rate of  $10 \text{ L min}^{-1}$  (room temperature and one atmospheric pressure) adjusted using a Cole-Palmer<sup>®</sup> flowmeter (accuracy  $\pm 3\%$ ). In order to capture the daily variability in  $\text{NO}_2$  isotopic composition, samples were collected during 24 h with 3 h sampling intervals during the day and 5 h sampling from midnight to 05:00

local time (LT). Ambient NO and NO<sub>2</sub> mixing ratios were measured with a 2B Technologies® NO monitor (model 410) paired with a NO<sub>2</sub> converter (model 401).

Honeycomb denuders were cleaned and coated the day before sampling. After being generously rinsed (5 min under a stream of deionized water), the denuders were placed in a vacuum chamber (Thermo Scientific® refrigerated vapour trap paired with a SpeedVac concentrator) and dried at 40 °C during 1 h. Then, denuder's internal walls were individually coated with 10 mL of a 95 : 5 by volume mixture of 2.5 M KOH (in methanol) and ultrapure guaiacol prepared daily. Denuders were then drawn off, dried in the vacuum chamber at 40 °C during 30 min to minimize blanks, hermetically sealed, and stored at ambient temperature in the dark until usage. The different components of the cartridge (impactor, filters, denuders) were cleaned, dried, and fitted together just before use. At the end of the sampling period, both denuders were removed from the ChemComb cartridge and rinsed with 10 mL of deionized water in order to leach trapped NO<sub>2</sub> out. A volume of 1 mL of the eluent was rapidly used to determine the nitrite amount using the Griess–Saltzman reaction and UV–Vis spectrometry at 544 nm. Recovered eluent ( $\approx 7$  mL by denuder) was poured into a labelled 15 mL Corning® tube and stored in a freezer until isotopic analysis the following days.

### 3 Atmospheric observations and multi-isotopic measurements

#### 3.1 NO<sub>x</sub> and O<sub>3</sub> atmospheric observations

Figure 2 shows the time evolution of the hourly NO<sub>2</sub>, NO, and O<sub>3</sub> mixing ratios measured during the period covering two nights and 1 d (from 14 May 2019 21:00 LT to 16 May 2019 05:00 LT). Note that most of our NO measurements are found to be within the reported detection limit of the instrument except in the morning (see Table 1) and therefore have to be treated with a lot of caution. NO<sub>2</sub> mixing ratios during the sampling period ( $(6.1 \pm 4.2)$  nmol mol<sup>-1</sup>; mean  $\pm 1$  standard deviation) are in good agreement with the range of values measured at the local air quality site located a kilometre south of the sampling site (<https://www.atmo-auvergnerhonealpes.fr/>, last access: 18 August 2020).

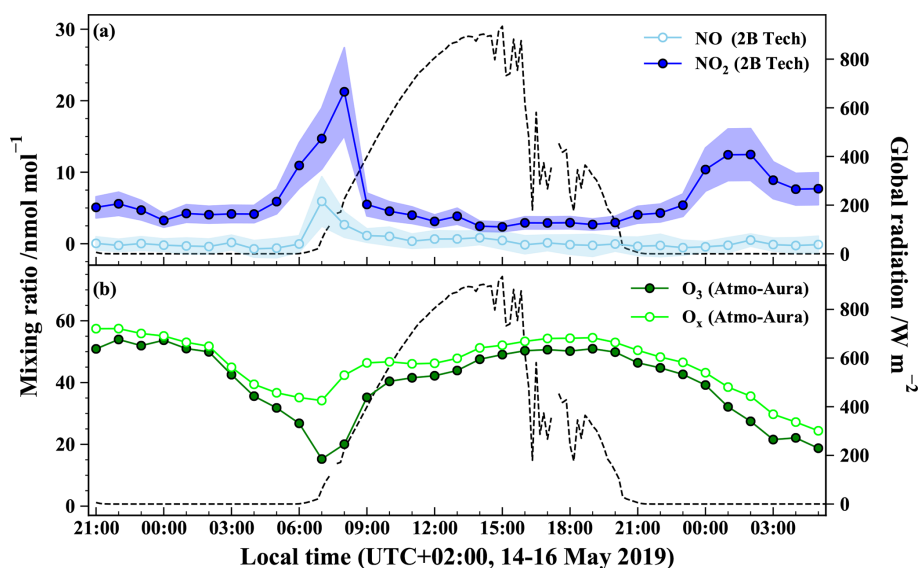
During both nights, most of the NO<sub>x</sub> is in the form of NO<sub>2</sub>. After sunrise, there is a rapid interconversion between NO and NO<sub>2</sub>, driven by NO<sub>2</sub> photolysis and reactions of NO with O<sub>3</sub> and peroxy radicals (Jacob, 1999). NO<sub>2</sub> levels are at a maximum on 15 May between 04:00 and 10:00 LT with a sharp peak of 21 nmol mol<sup>-1</sup> at 08:00 LT. After the morning rise, NO<sub>2</sub> decreases to reach a background mixing ratio of about  $(3.0 \pm 0.5)$  nmol mol<sup>-1</sup>. This diurnal variation is common in urban/suburban sites characterized by a morning peak caused by important NO<sub>x</sub> emissions, mainly from road traffic (Mayer, 1999). As morning progresses, the boundary

layer height increases rapidly, favouring fast dilution of NO<sub>x</sub> mixing ratios. Moreover, during the day, NO<sub>2</sub> is converted to HNO<sub>3</sub>, notably by its reaction with OH radicals. Thus, NO<sub>x</sub> mixing ratio remains low during the day likely because of the combination of atmospheric dilution by vertical mixing and efficient chemical conversion by OH and organic radicals (Tie et al., 2007). In dense urban areas, a second NO<sub>x</sub> traffic emission peak can occur in late afternoon, but it is not observed at our sampling site for that specific day. This surface pollution peak is usually weaker than the morning peak due to an elevated boundary layer and a longer period of evening commuting. After sunset, NO<sub>2</sub> mixing ratio increases gently and reaches a smooth peak with a maximum of 12 nmol mol<sup>-1</sup> around 01:00 LT, also recorded at the local air quality site. This NO<sub>2</sub> mixing ratio rise may be due to low NO emissions (converted to NO<sub>2</sub> by reaction with O<sub>3</sub>) combined with a decreasing boundary layer height during the night which traps atmospheric species close to the surface (Tie et al., 2007; Villena et al., 2011).

Ozone also exhibits a diurnal variation typical of urban areas (Velasco et al., 2008). O<sub>3</sub> peaks around 50 nmol mol<sup>-1</sup> at the beginning of both nights and then declines continuously. Indeed, after sunset, O<sub>3</sub> production ceases, and its mixing ratio drops due to its dry deposition, reactions with organics, and O<sub>3</sub> titration by NO emitted from evening traffic, and industrial activities in the stable nocturnal boundary layer (Klein et al., 2019). O<sub>3</sub> reaches a minimum (about 15 nmol mol<sup>-1</sup>) not at the end of the night but during the morning rush hour peak of NO. The O<sub>x</sub> (= O<sub>3</sub> + NO<sub>2</sub>) is a more conservative quantity than O<sub>3</sub>, because it is less affected by conversion of O<sub>3</sub> into NO<sub>2</sub> through NO titration which is important in urban environments (Kleinman et al., 2002). For instance, between 06:00 and 08:00 LT, O<sub>3</sub> is strongly titrated by freshly emitted NO with its mixing ratio dropping to about 15 nmol mol<sup>-1</sup>, while O<sub>x</sub> reaches a moderate minimum of 34 nmol mol<sup>-1</sup>. After this morning drop, O<sub>3</sub> recovers rapidly to about 46 nmol mol<sup>-1</sup> in the late morning, possibly caused by downward O<sub>3</sub> flux associated with the formation of the daytime thick boundary layer (Jin and Demerjian, 1993; Klein et al., 2019). During the rest of the day, O<sub>3</sub> and O<sub>x</sub> keep increasing gently due to photochemical production and reach a close maximum at the end of the afternoon (Geng et al., 2008). After sunset, the important decline of both O<sub>3</sub> and O<sub>x</sub> highlights the physical losses, notably O<sub>3</sub> deposition, and chemical loss of NO<sub>x</sub>, typical of urban areas.

#### 3.2 Multi-isotopic composition measurements of atmospheric NO<sub>2</sub>

We present the data for the multi-isotopic composition of seven atmospheric NO<sub>2</sub> samples, while two additional samples were rejected as NO<sub>2</sub><sup>-</sup> amounts were too low to perform a reliable analysis. Table 1 reports ambient mean mixing ratios of NO, NO<sub>2</sub>, and O<sub>3</sub> for the isotopic sampling intervals and corresponding measured NO<sub>2</sub> isotopic composition



**Figure 2.** Temporal evolution of (a) NO (open circles) and NO<sub>2</sub> (closed circles) at the sampling site (the envelopes represent  $\pm 1\sigma$  variations over 1 h) and of (b) O<sub>3</sub> (closed circles) and O<sub>x</sub> (= O<sub>3</sub> + NO<sub>2</sub>; open circles) at the air quality station during the sampling period. Markers represent for (a) the hourly mean derived from 1 min measurements and for (b) the hourly mean provided by the air quality monitoring station. Global solar radiation flux is represented by dashed lines (measured at 200 m from the sampling site by the IGE weather station with a Skye SP1110 pyranometer).

**Table 1.** Summary table of sampling periods (dates, local times), NO, NO<sub>2</sub>, and O<sub>3</sub> mean mixing ratios over the collection periods, as well as calibrated NO<sub>2</sub> isotopic measurements of  $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$ . All the sampling periods lasted 3 h except the last one that lasted 5 h. Averaged measurement uncertainties are provided just below the species names.

Local sampling date & time (start–end)	NO ( $\pm 2.5 \text{ nmol mol}^{-1}$ )	NO <sub>2</sub> ( $\pm 2.5 \text{ nmol mol}^{-1}$ )	O <sub>3</sub> <sup>*</sup> ( $\pm 6.8 \text{ nmol mol}^{-1}$ )	$\delta^{15}\text{N}(\text{NO}_2)$ ( $\pm 0.1\text{‰}$ )	$\delta^{18}\text{O}(\text{NO}_2)$ ( $\pm 1.8\text{‰}$ )	$\Delta^{17}\text{O}(\text{NO}_2)$ ( $\pm 0.3\text{‰}$ )
14 May 2019 21:00–00:00	0.0	5.1	52.3	−11.7	75.6	27.4
15 May 2019 06:00–09:00	2.9	15.6	20.7	−4.9	97.6	31.8
15 May 2019 09:00–12:00	0.8	4.7	39.1	−10.1	114.5	39.2
15 May 2019 12:00–15:00	0.7	3.1	44.6	−11.8	90.9	35.8
15 May 2019 15:00–18:00	0.2	2.7	50.0	−11.0	86.9	31.1
15 May 2019 18:00–21:00	0.0	2.9	50.3	−11.1	77.1	29.7
16 May 2019 00:00–05:00	0.0	9.9	26.9	−11.1	62.2	20.5

\* Data monitored at the local air quality monitoring site of Saint-Martin d'Hères located a kilometre south of the sampling site (<https://www.atmo-auvergnerhonealpes.fr/>).

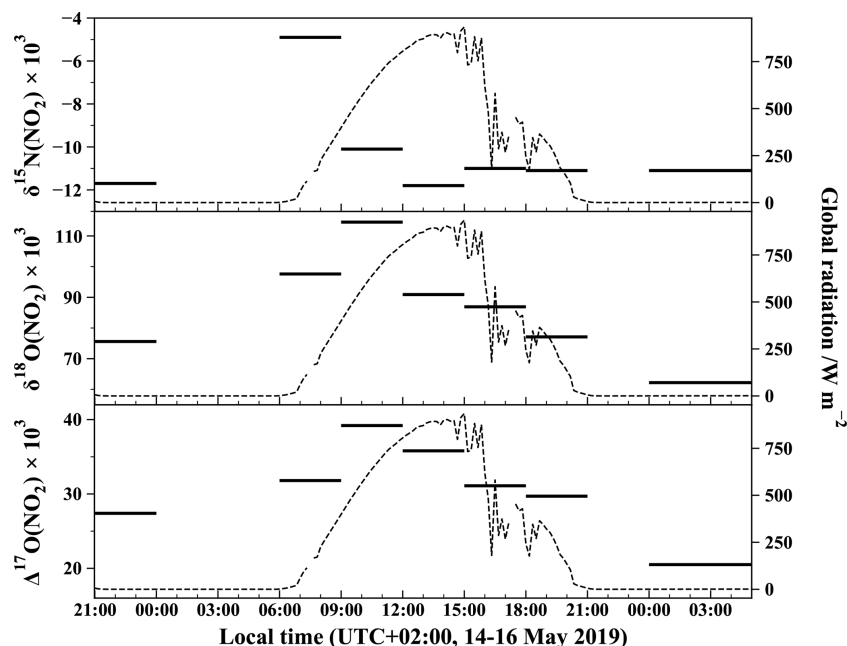
( $\delta^{15}\text{N}(\text{NO}_2)$ ,  $\delta^{18}\text{O}(\text{NO}_2)$ , and  $\Delta^{17}\text{O}(\text{NO}_2)$ ). Figure 3 depicts the time series of measured  $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$  of atmospheric NO<sub>2</sub>. The temporal evolution of NO<sub>2</sub> oxygen and nitrogen isotopic composition is interpreted in the following section.

## 4 Discussion of the multi-isotopic composition of atmospheric NO<sub>2</sub>

### 4.1 Oxygen isotope composition

The time evolution of  $\delta^{18}\text{O}$  of atmospheric NO<sub>2</sub> ( $\delta^{18}\text{O}(\text{NO}_2)$ ) shown in Fig. 3 exhibits a substantial diurnal variation with a daytime mean of  $(93.4 \pm 13.9)\text{‰}$

and a nighttime mean of  $(68.9 \pm 9.5)\text{‰}$ . A maximum value of 114.5‰ is observed in the morning (09:00–12:00 LT interval) and a minimum value of 62.2‰ for the late-night interval (00:00–05:00 LT). Using a similar sampling apparatus during summer in the urban/suburban site of West Lafayette, USA, Walters et al. (2018) reported daytime and nighttime mean  $\delta^{18}\text{O}(\text{NO}_2)$  values of  $(86.5 \pm 14.1)\text{‰}$  and  $(56.3 \pm 7.1)\text{‰}$ , respectively. Although our daytime values are higher than those of Walters et al. (2018), both datasets exhibit the same day–night contrast with a maximum during the day and a minimum at night. As expected from  $\delta^{18}\text{O}$  values,  $\Delta^{17}\text{O}(\text{NO}_2)$  follows a similar diurnal variation with a maximum value of 39.2‰ for the 09:00–12:00 LT interval and a minimum value of 20.5‰ for

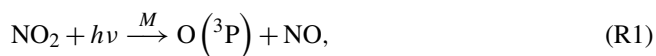


**Figure 3.** Temporal evolution of  $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$  of atmospheric  $\text{NO}_2$  measured with the azide method. Isotopic values for each 3 h slots are from the same  $\text{NO}_2$  sample collected over 3 h (except for the last period which lasted 5 h). Global solar radiation flux is represented by dashed lines (measured at 200 m from the sampling site by the IGE weather station with a Skye SP1110 pyranometer).

the 00:00–05:00 LT interval. High  $\Delta^{17}\text{O}$  values are expected to reflect the importance of ozone in the oxidation of NO to  $\text{NO}_2$ . Since daytime and nighttime chemistries are radically different, interpretations of our  $\Delta^{17}\text{O}$  measurements and their implications are discussed separately by day and night.

#### 4.1.1 Fundamentals of $\text{NO}_x$ chemistry and isotopic transfers

$\text{NO}_x$  is mainly produced under the form of NO by combustion and lightning processes (Dennison et al., 2006; Young, 2002) as well as by the biological activity of soils (Davidson and Kingerlee, 1997). In the daytime, NO and  $\text{NO}_2$  rapidly interconvert in a timescale of about 1–2 min, establishing a photostationary steady state (PSS; Leighton 1961):



with  $M = \text{N}_2$  or  $\text{O}_2$ ;



This so-called null cycle can be disturbed by  $\text{RO}_2$  radicals when  $\text{NO}_x$  mixing ratios are relatively high, typically above 30–100  $\text{pmol mol}^{-1}$  (Seinfeld and Pandis, 2006; Monks, 2005):

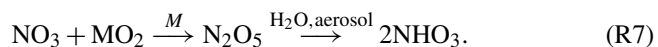


The reaction between NO and  $\text{RO}_2$  competes with the  $\text{NO} + \text{O}_3$  reaction, allowing  $\text{NO}_2$  formation without the consumption of an ozone molecule in the cycle (Monks, 2005). This results in ozone production and can lead to severe ozone build-up in polluted areas. At night,  $\text{RO}_2$  mixing ratios are strongly reduced making ozone the main NO oxidant following Reaction (R3).

$\text{NO}_x$  is mainly removed from the atmosphere via the oxidation of  $\text{NO}_2$  into nitric acid during the day,



and at night,



In this framework,  $\Delta^{17}\text{O}(\text{NO}_2)$  is driven by the relative importance of the different  $\text{NO}_2$  production channels, because  $\text{NO}_2$  loss processes do not fractionate in terms of oxygen mass-independent anomaly. Each  $\text{NO}_2$  production channel generates a specific mass-independent isotopic anomaly  $\Delta^{17}\text{O}$  on the produced  $\text{NO}_2$  (Kaiser et al., 2004). Based on the  $\text{NO}_2$  continuity equation, this can be expressed with the following  $\Delta^{17}\text{O}(\text{NO}_2)$  mass-balance equation (Morin et al., 2011):

$$\frac{d}{dt} \left( [\text{NO}_2] \times \Delta^{17}\text{O}(\text{NO}_2) \right) = \sum_i \left( P_i \times \Delta^{17}\text{O}_i(\text{NO}_2) \right) - \left( \sum_j L_j \right) \times \Delta^{17}\text{O}(\text{NO}_2), \quad (2)$$

with  $[\text{NO}_2]$  being the atmospheric  $\text{NO}_2$  mixing ratio,  $P_i$  and  $L_j$  the  $\text{NO}_2$  production/emission and loss rates, respectively (i.e. mixing ratio of involved species multiplied by the kinetic constant of the considered chemical reaction), and  $\Delta^{17}\text{O}_i(\text{NO}_2)$  the specific isotope anomaly transferred to  $\text{NO}_2$  through the production reaction  $i$ .

#### 4.1.2 $\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)$

By day, the  $\text{NO}_x$  photochemical cycle (Reactions R1 to R4) achieves a steady state in 1–2 min, which is several orders of magnitude faster than  $\text{NO}_2$  loss reactions (Atkinson et al., 1997) and emission rate ( $\text{NO}_x$  is mainly emitted under the form of  $\text{NO}$ ; Villena et al., 2011). It follows that  $\text{NO}$  and  $\text{NO}_2$  short variations can be neglected; i.e.  $\frac{d}{dt}[\text{NO}_2] \approx 0$  and  $\frac{d}{dt}[\text{NO}] \approx 0$  on short timescales. In addition, fast interconversions between  $\text{NO}$  and  $\text{NO}_2$  generate quickly an isotopic equilibrium between  $\text{NO}$  and  $\text{NO}_2$  resulting in  $\Delta^{17}\text{O}(\text{NO}_2) \approx \Delta^{17}\text{O}(\text{NO})$  (Michalski et al., 2014; Morin et al., 2007). With these approximations, considering only the main reactions and neglecting halogen chemistry, Eq. (2) yields the following (Morin et al., 2007):

$$\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2) \approx \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3] \times \Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) + k_{\text{NO}+\text{RO}_2}[\text{RO}_2] \times \Delta^{17}\text{O}_{\text{NO}+\text{RO}_2}(\text{NO}_2)}{k_{\text{NO}+\text{O}_3}[\text{O}_3] + k_{\text{NO}+\text{RO}_2}[\text{RO}_2]}, \quad (3)$$

with  $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$  being the ozone isotopic anomaly transferred to  $\text{NO}$  during its oxidation to  $\text{NO}_2$  via Reaction (R3) (also called the transfer function of the isotope anomaly of ozone to  $\text{NO}_2$ ; Savarino et al., 2008) and  $\Delta^{17}\text{O}_{\text{NO}+\text{RO}_2}(\text{NO}_2)$  being the  $\text{RO}_2$  isotopic anomaly transferred to  $\text{NO}$  during its oxidation to  $\text{NO}_2$  via Reaction (R4).  $\Delta^{17}\text{O}_{\text{NO}+\text{RO}_2}(\text{NO}_2)$  can be considered to be negligible (Alexander et al., 2020; Michalski et al., 2003), because  $\text{RO}_2$  is mainly formed by the reactions  $\text{R} + \text{O}_2$  and  $\text{H} + \text{O}_2$  and because the isotopic anomaly of atmospheric  $\text{O}_2$  is very close to 0‰ (Barkan and Luz, 2003). This assumption has been estimated to affect the overall  $\Delta^{17}\text{O}$  of  $\text{RO}_2$  values by less than 1‰ (Röckmann et al., 2001). As a result, Eq. (3) can be simplified, giving a  $\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)$  driven by the relative importance of Reaction (R3) ( $\text{NO} + \text{O}_3$ ) and Reaction (R4) ( $\text{NO} + \text{RO}_2$ ) in the  $\text{NO}$  oxidation and by the oxygen isotopic anomaly transferred from  $\text{O}_3$  to  $\text{NO}_2$ :

$$\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2) \approx T_{\text{NO}+\text{O}_3} \times \Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2), \quad (4)$$

with

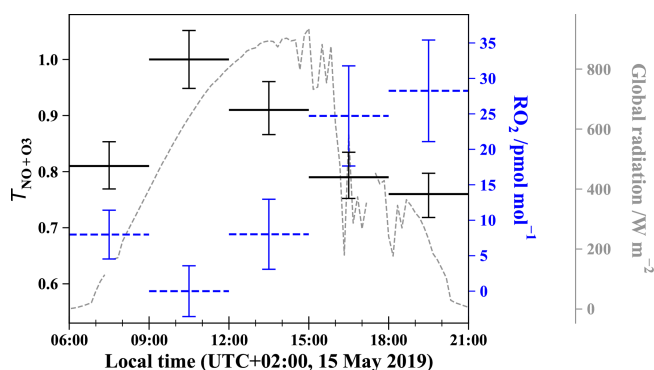
$$T_{\text{NO}+\text{O}_3} = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3]}{k_{\text{NO}+\text{O}_3}[\text{O}_3] + k_{\text{NO}+\text{RO}_2}[\text{RO}_2]}. \quad (5)$$

$\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$  has been determined experimentally by Savarino et al. (2008). They reported  $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) = 1.18 \pm 0.07 \times \Delta^{17}\text{O}(\text{O}_3) + (6.6 \pm 1.5)\text{‰}$  with  $\Delta^{17}\text{O}(\text{O}_3)$  being the bulk ozone isotopic anomaly.  $\Delta^{17}\text{O}(\text{O}_3)$  has been measured in Grenoble in 2012 (Vicars and Savarino, 2014) with a mean value of  $(26.2 \pm 1.3)\text{‰}$ , corresponding to a  $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$  value of  $(37.5 \pm 2.8)\text{‰}$ , which, according to Eq. (4), would give a maximum  $\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)$  value of  $(37.5 \pm 2.8)\text{‰}$ . It is consistent with our maximum measured  $\Delta^{17}\text{O}(\text{NO}_2)$  value of 39.2‰ for the 09:00–12:00 LT interval. In light of the known uncertainties, the small difference is not significant and is much smaller than the diurnal variations of  $\Delta^{17}\text{O}(\text{NO}_2)$ . Note that the  $\Delta^{17}\text{O}$  calibration is not very accurate for the most enriched samples, because nitrite standards with high  $\Delta^{17}\text{O}$  are still not readily available. In a laboratory study, Michalski et al. (2014) measured the  $\Delta^{17}\text{O}$  of  $\text{NO}_2$  formed by the photochemical  $\text{NO}-\text{NO}_2-\text{O}_3$  cycle and reported  $\Delta^{17}\text{O}(\text{NO}_2) = (39.3 \pm 1.9)\text{‰}$ . Despite experimental conditions that are not strictly applicable to our atmospheric conditions (e.g.  $\text{NO}_x \gg \text{O}_3$ , light source, absence of volatile organic compounds (VOCs)), their value is surprisingly close to our maximum value. Assuming that our maximum  $\Delta^{17}\text{O}(\text{NO}_2)$  value corresponds to  $T_{\text{NO}+\text{O}_3}$  close to unity (Reaction R3 ( $\text{NO} + \text{O}_3$ )  $\gg$  Reaction R4 ( $\text{NO} + \text{RO}_2$ )), we use a  $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$  value of 39.2‰ for the following calculations. Combining Eqs. (4) and (5), an expression for the  $\text{RO}_2$  mixing ratio can be derived as

$$[\text{RO}_2] = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3]}{k_{\text{NO}+\text{RO}_2}} \left( \frac{\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)}{\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)} - 1 \right). \quad (6)$$

Figure 4 shows the daytime evolution of  $T_{\text{NO}+\text{O}_3}$  calculated from Eq. (4) and  $\text{RO}_2$  calculated from Eq. (6).  $T_{\text{NO}+\text{O}_3}$  varies between 0.76 and 1 with a mean daytime value of 0.86 (the measured daytime  $\Delta^{17}\text{O}(\text{NO}_2)$  mean value is  $(33.5 \pm 3.9)\text{‰}$ ), meaning that 86 % of  $\text{NO}_2$  is formed via Reaction (R3) (oxidation of  $\text{NO}$  by  $\text{O}_3$ ). The mean estimated  $\text{RO}_2$  mixing ratio is  $(13.8 \pm 11.2)\text{ pmol mol}^{-1}$ . Note that  $\text{RO}_2 = 0\text{ pmol mol}^{-1}$  for the 09:00–12:00 LT interval originates from our assumption of  $T_{\text{NO}+\text{O}_3} = 1$  for our highest  $\Delta^{17}\text{O}(\text{NO}_2)$  value; in reality, it only means that  $\text{RO}_2$  is so low that Reaction (R3) ( $\text{NO} + \text{O}_3$ )  $\gg$  Reaction (R4) ( $\text{NO} + \text{RO}_2$ ). Overall, our  $\text{RO}_2$  values are found to be within the range of values measured at urban/peri-urban sites (see Table 2). However,  $\text{RO}_2$  diurnal variation at our site does not follow the pattern of previous measurements which usually report a diurnal variation with a maximum varying from noon to early afternoon (Fuchs et al., 2008; Tan et al., 2017), whereas this study shows a maximal mixing ratio in late afternoon. With such a limited dataset (only 1 d of sampling), it is not possible to draw general conclusions on the  $\text{NO}_x$  /  $\text{RO}_2$  chem-





**Figure 4.** Daytime evolution of  $T_{\text{NO}+\text{O}_3}$  (solid black lines), estimated from Eq. (4) using measured  $\Delta^{17}\text{O}(\text{NO}_2)$  in Grenoble, and of  $\text{RO}_2$  mixing ratios (dashed blue lines), estimated from Eq. (6). Error bars for  $T_{\text{NO}+\text{O}_3}$  are derived from standard deviations of  $\Delta^{17}\text{O}(\text{NO}_2)$  and  $\Delta^{17}\text{O}(\text{O}_3^*)$  measured in Grenoble (Vicars and Savarino, 2014).  $\text{RO}_2$  error bars are derived from  $\text{O}_3$  measurement uncertainties and errors on  $T_{\text{NO}+\text{O}_3}$  (by comparison, errors on reaction constants can be neglected). Global solar radiation flux is represented by dashed lines (measured at 200 m from the sampling site by the IGE weather station with a Skye SP1110 pyranometer).

istry dynamics. An important recommendation for further investigation is to conduct isotopic measurements with accurate measurements of key atmospheric radicals/oxidants, e.g. NO,  $\text{O}_3$ , and possibly  $\text{RO}_2$ , in order to test quantitatively our isotopic approach. Additionally, the use of a chemical box model is also recommended, because it would allow us to account for non-equilibrium effects in isotopic transfers and thus strengthen the interpretation of isotopic measurements in the investigation of the reactive nitrogen cycle in urban atmospheres.

Morin et al. (2011) simulated the diurnal variation of  $\Delta^{17}\text{O}(\text{NO}_2)$  in a remote marine boundary layer without the effect of  $\text{NO}_x$  emissions. They assumed  $\Delta^{17}\text{O}(\text{O}_3) = 30\text{‰}$  ( $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2) = 45\text{‰}$ ) resulting in higher overall  $\Delta^{17}\text{O}(\text{NO}_2)$  values compared to our study. Their simulated  $\Delta^{17}\text{O}(\text{NO}_2)$  exhibited large diurnal variations with maximum values at night (close to  $41\text{‰}$ ) and minimum values at noon of  $28\text{‰}$ . This is consistent with  $\text{RO}_2$  mixing ratio reaching a maximum around local noon in clean environments. In contrast to their model simulations, our daytime  $\Delta^{17}\text{O}(\text{NO}_2)$  measurements are higher than our nighttime measurements. We will show in the following section that this difference originates from the absence of  $\text{NO}_x$  emissions in the Morin et al. (2011) photochemical modelling.

#### 4.1.3 $\Delta^{17}\text{O}_{\text{night}}(\text{NO}_2)$

Without photolysis at night and associated  $\text{RO}_2$  production, ozone is the unique NO oxidant. NO and  $\text{NO}_2$  are no longer in photochemical equilibrium, because  $\text{NO}_2$  cannot be converted back into NO. As a result, the oxygen isotopic compo-

sition of  $\text{NO}_2$  formed during the night is determined by the oxygen isotopic composition of  $\text{O}_3$  and emitted NO. Additionally, in order to estimate the overall isotopic signature of sampled  $\text{NO}_2$  at night, we need to determine the residuals of  $\text{NO}_2$  formed during the day that are still present during the night:

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

with  $x$  being the fraction of  $\text{NO}_2$  formed during the day to the total  $\text{NO}_2$  measured at night and with  $(1-x)$  representing the fraction of  $\text{NO}_2$  which has been produced during the night to the total  $\text{NO}_2$  measured at night. NO is mainly emitted by combustion processes in which a nitrogen atom (from atmospheric  $\text{N}_2$  or N present in fuel) is added to an oxygen atom formed by the thermal decomposition of  $\text{O}_2$  (Zeldovich, 1946). With  $\Delta^{17}\text{O}(\text{O}_2)$  being close to  $0\text{‰}$  (Barkan and Luz, 2003), NO emissions are very likely to have a  $\Delta^{17}\text{O} \approx 0\text{‰}$  or at least be negligible compared to  $\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ . Using Eq. (7) and assuming a negligible isotope anomaly for NO, the time evolution of  $\Delta^{17}\text{O}(\text{NO}_2)$  over the night can be calculated. It is worth pointing out that the  $x$  fraction becomes very small at the end of the night, allowing us to further simplify Eq. (7):  $\Delta^{17}\text{O}_{\text{night}}(\text{NO}_2) = \frac{1}{2} \times \Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)$ . Thus, if there are nighttime NO emissions, a measurement of  $\Delta^{17}\text{O}(\text{NO}_2)$  at the end of the night is also an interesting way of deriving  $\Delta^{17}\text{O}(\text{O}_3)$  which is difficult to measure directly. The nighttime variation of the  $x$  fraction is estimated considering that the nighttime lifetime of  $\text{NO}_2$  relative to oxidation via ozone and dry deposition is 7.2 h ( $\text{O}_3$  chemical sink is dominant over deposition by a factor  $> 10^4$  with  $k_{\text{NO}_2+\text{O}_3} = 1.4 \times 10^{-13} \exp[-2470(K)/T] \text{ cm}^3 \text{ mol}^{-1} \text{ s}^{-1}$ ; Atkinson et al., 2004;  $\text{NO}_2$  dry velocity  $V_d = 0.25 \text{ cm s}^{-1}$ ; Holland et al., 1999; and assuming a nighttime boundary layer height of 500 m). For the 00:00–05:00 LT interval, we calculate a mean value of  $\Delta^{17}\text{O}(\text{NO}_2) = 19.9\text{‰}$  (with an overall error of about  $1.6\text{‰}$ ), which is very close to our measured  $\Delta^{17}\text{O}(\text{NO}_2)$  of  $20.5\text{‰}$ . This first dataset of nighttime  $\Delta^{17}\text{O}(\text{NO}_2)$  measurements appears to confirm our understanding of nocturnal  $\text{NO}_2$  formation (Alexander et al., 2020; Michalski et al., 2014). NO emissions in urban areas have a very significant influence on  $\Delta^{17}\text{O}(\text{NO}_2)$ , leading to a behaviour in opposition to the one observed in remote locations. As illustrated by Morin et al. (2011),  $\Delta^{17}\text{O}(\text{NO}_2)$  is predicted to be maximal at night in remote areas where NO emissions are negligible, reflecting the isotopic signature of  $\text{NO}_2$  at sunset. In areas where nighttime NO emissions are high, nighttime  $\Delta^{17}\text{O}(\text{NO}_2)$  can be up to a factor of 2 smaller than in remote areas.



**Table 2.** Mean daytime RO<sub>2</sub> mixing ratio ranges measured during field campaigns in various environments and seasons.

Site	RO <sub>2</sub> (pmol mol <sup>-1</sup> )	Reference
Grenoble (2019, May)	0–35*	This study
UK, suburban site (2003, July–August)	4–22	Emmerson et al. (2007)
Germany, suburban site (2005, July)	2–40	Fuchs et al. (2008)
Germany, rural site (1998, July–August)	2–50	Mihelcic et al. (2003)
USA, rural site (2002, May–June)	9–15	Ren et al. (2005)
China, rural site (2014, June–July)	7–37	Tan et al. (2017)

\* Derived from Eq. (6) using  $\Delta^{17}\text{O}$  values of atmospheric NO<sub>2</sub> in Grenoble.

## 4.2 Nitrogen isotope composition

Measured  $\delta^{15}\text{N}(\text{NO}_2)$  values range from  $-11.8\text{‰}$  to  $-4.9\text{‰}$  with no clear diurnal variation and values clustering around an overall mean of  $(-10.2 \pm 2.2)\text{‰}$  (see Fig. 3). Using a similar method, Walters et al. (2018) collected atmospheric NO<sub>2</sub> over 1 month in a urban/suburban location during the summer. They reported a mean  $\delta^{15}\text{N}$  value of  $(-11.4 \pm 6.9)\text{‰}$ , very close to our mean value but with a wider overall range (from  $-31.4\text{‰}$  to  $+0.4\text{‰}$ ). In another urban area but using passive samplers, Dahal and Hastings (2016) reported mean  $\delta^{15}\text{N}(\text{NO}_2)$  values of  $(-8.3 \pm 0.9)\text{‰}$  and  $(-6.4 \pm 1.4)\text{‰}$  for summer and winter periods, respectively. All these values are within the  $\delta^{15}\text{N}$  range for NO emitted by industrial combustion and traffic sources which are reported to vary from  $-19.7\text{‰}$  to  $-13.7\text{‰}$  and from  $-9\text{‰}$  to  $-2\text{‰}$ , respectively (Miller et al., 2017; Walters et al., 2015). Interestingly, all the  $\delta^{15}\text{N}$  values measured at our sampling site fall within a narrow range, from about  $-12\text{‰}$  to  $-10\text{‰}$ , except for the sample collected between 06:00 and 09:00 LT which has a much higher value of  $-4.9\text{‰}$ . This singular value is well correlated with the morning NO traffic emission spike (see Fig. 2). However, once emitted into the atmosphere, NO can undergo isotopic fractionations that modify the nitrogen isotope distribution in NO<sub>2</sub> relative to emitted NO (Freyer et al., 1993). In order to use  $\delta^{15}\text{N}(\text{NO}_2)$  as a tracer of NO<sub>x</sub> sources, we need to quantify these nitrogen isotopic shifts to correct measured  $\delta^{15}\text{N}(\text{NO}_2)$ . Nitrogen isotopic fractionation, defined as  $\Delta(\text{NO}_2 - \text{NO}_x) = \delta^{15}\text{N}(\text{NO}_2) - \delta^{15}\text{N}(\text{NO}_x)$ , is the result of a combination of three effects: (1) an equilibrium isotope effect (EIE) between NO and NO<sub>2</sub>, (2) a kinetic isotope effect (KIE) during NO oxidation to NO<sub>2</sub>, and (3) a photochemical isotope fractionation effect (PHIFE) during NO<sub>2</sub> photolysis (other NO<sub>2</sub> sinks are negligible during the day). The overall daytime nitrogen isotopic shift of NO<sub>2</sub> relative to emitted NO<sub>x</sub> ( $\Delta_{\text{day}}(\text{NO}_2 - \text{NO}_x)$ ) can be estimated using the steady-state isotopic mass balance for NO<sub>2</sub>. Li et al. (2020) derived an expression for  $\Delta(\text{NO}_2 - \text{NO}_x)$  by assuming that the conversion of NO to NO<sub>2</sub> is solely driven by O<sub>3</sub>. This could therefore lead to uncertainties on the NO<sub>2</sub> shift when

other conversion pathways become significant with respect to the NO conversion by O<sub>3</sub>. A more general expression for  $\Delta(\text{NO}_2 - \text{NO}_x)$  can be derived by taking into account the conversion of NO to NO<sub>2</sub> via other species, notably RO<sub>2</sub> (see Eq. C11 in Appendix C). In our urban environment, we only consider the conversion of NO into NO<sub>2</sub> via O<sub>3</sub> and RO<sub>2</sub> during the day. Assuming  $\alpha_{\text{KIE}(\text{NO}+\text{O}_3)} \approx \alpha_{\text{KIE}(\text{NO}+\text{RO}_2)}$  (see derivation in Appendix C),  $\Delta_{\text{day}}(\text{NO}_2 - \text{NO}_x)$  can be expressed by

$$\Delta_{\text{day}}(\text{NO}_2 - \text{NO}_x) = \frac{\alpha_{\text{LCIE}}^* A_{\text{day}}^* + (\alpha_{\text{EIE}} - 1)}{A_{\text{day}}^* + 1} \times (1 - f_{\text{NO}_2}), \quad (8)$$

with

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

and

$$A_{\text{day}}^* = \frac{J_{\text{NO}_2}}{k_{\text{NO}+\text{NO}_2}[\text{NO}]} = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3] + k_{\text{NO}+\text{RO}_2}[\text{RO}_2]}{k_{\text{NO}+\text{NO}_2}[\text{NO}_2]},$$

where  $f_{\text{NO}_2} = [\text{NO}_2]/[\text{NO}_x]$ ,  $\alpha_{\text{LCIE}}^*$  is the fractionation factor of combined KIE and PHIFE (LCIE represents the Leighton cycle isotope effect), and  $\alpha_{\text{EIE}}$  is the EIE fractionation factor.  $A_{\text{day}}^*$  is defined as the ratio of the NO<sub>2</sub> lifetime with respect to isotopic exchanges over the daytime NO<sub>2</sub> chemistry lifetime.  $J_{\text{NO}_2}$  is the NO<sub>2</sub> photolysis rate,  $k_{\text{NO}+\text{O}_3}$  is the rate constant of reaction NO + O<sub>3</sub>,  $k_{\text{NO}+\text{RO}_2}$  is the rate constant of reaction NO + RO<sub>2</sub>, and  $k_{\text{NO}+\text{NO}_2}$  is the rate constant of the isotopic exchange (Reaction CR1) (see Appendix D for rate constants data). Interestingly, we can combine Eqs. (8) and (6) and express  $A_{\text{day}}^*$  as a function of oxygen isotopic variables discussed in the previous section:

$$A_{\text{day}}^* = \frac{k_{\text{NO}+\text{O}_3}[\text{O}_3]}{k_{\text{NO}+\text{NO}_2}[\text{NO}_2]} \left( \frac{\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)}{\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)} \right). \quad (9)$$

Since our NO measurements are not precise, and we do not have direct measurements of  $J_{\text{NO}_2}$  or RO<sub>2</sub>, we use Eq. (9) to estimate the NO<sub>2</sub> isotopic fractionation shift. Note that although Li et al. (2020) only consider the NO conversion via

O<sub>3</sub> in the analysis of their nitrogen isotopic data, they found an excellent agreement between their calculated values and field isotopic measurements at Jülich, Germany (Freyer et al., 1993). Nonetheless, the reason for this accordance remains unclear, as it could be attributable to an equivalent KIE of NO + O<sub>3</sub> and NO + RO<sub>2</sub> but also to the dominance of the NO oxidation via O<sub>3</sub> over RO<sub>2</sub>.

At night, the isotopic fractionation shift  $\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x)$  is driven by EIE, KIE, the N isotopic composition of NO emissions and  $f_{\text{NO}_2}$ , given that  $J_{\text{NO}_2}$  is null (see derivation in Appendix C):

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x) \approx \frac{A_{\text{night}}^* \left( \alpha_{\text{KIE}} - \left( \frac{1 + \delta^{15}\text{N}(\text{NO}_{\text{emis}})}{1 + \delta^{15}\text{N}(\text{NO}_2)} \right) \right) + (\alpha_{\text{EIE}} - 1)}{A_{\text{night}}^* + 1} (1 - f_{\text{NO}_2}), \quad (10)$$

where  $\delta^{15}\text{N}(\text{NO}_{\text{emis}})$  is the N isotopic composition of NO emissions, and

$$A_{\text{night}}^* = \frac{k_{\text{NO}+\text{O}_3} [\text{O}_3]}{k_{\text{NO}+\text{NO}_2} [\text{NO}_2]} = \frac{E(\text{NO})}{k_{\text{NO}+\text{NO}_2} [\text{NO}] [\text{NO}_2]}, \quad (11)$$

with  $E(\text{NO})$  the NO emission flux. From laboratory experiments, Li et al. (2020) reported fractionation factors of  $1.0289 \pm 0.0019$  and  $0.990 \pm 0.005$  for  $\alpha_{\text{EIE}}$  and  $\alpha_{\text{LCIE}}^*$ , respectively. Using these experimental values and the ambient mixing ratios of ozone, NO, and NO<sub>2</sub> measured at our sampling site, we estimate the time evolution of  $\Delta(\text{NO}_2 - \text{NO}_x)$  from Eqs. (8) and (9) for daytime. At night,  $[\text{NO}] \ll [\text{NO}_2]$ ; hence,  $f_{\text{NO}_2}$  tends towards 1 and  $\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x) \approx 0$  (Table 3 provides the calculated values).  $\Delta(\text{NO}_2 - \text{NO}_x)$  values are found to be negligible during the entire sampling period except between 06:00 and 09:00 LT with a mean  $\Delta(\text{NO}_2 - \text{NO}_x)$  value of 2.7‰ due to lower  $f_{\text{NO}_2}$  and  $A_{\text{day}}^*$  values. Overall, in our moderately polluted environment, nitrogen oxide isotope effects appear to induce very small nitrogen isotopic shift, reflecting the fact that NO<sub>x</sub> is overwhelmingly under the form of NO<sub>2</sub> (mean  $f_{\text{NO}_2} = 0.93$ ). Our results are in good agreement with the  $\Delta(\text{NO}_2 - \text{NO}_x)$  range (between 1.3 and 2.5‰) calculated from isotopic measurements at West Lafayette, USA (Walters et al., 2018). Moreover, Li et al. (2020) calculated a mean  $\Delta(\text{NO}_2 - \text{NO}_x)$  of  $(1.3 \pm 3.2)\text{‰}$  from isotopic measurements near San Diego, USA (NO<sub>x</sub> mixing ratios varied from 1 to 9 nmol mol<sup>-1</sup>).

Using estimated  $\delta^{15}\text{N}(\text{NO}_x)$ , we evaluate the relative contributions of the dominant NO<sub>x</sub> sources at our site using the Bayesian isotopic mixing model SIAR (Stable Isotope Analysis in R; Parnell et al., 2010). Initially developed for ecological studies (Inger et al., 2006; Samelius et al., 2007), isotopic mixing models have been recently used for atmospheric applications, notably to identify major NO<sub>x</sub> sources of aerosol nitrate from  $\delta^{15}\text{N}$  (Jin et al., 2021; Zong et al., 2017; Fan et al., 2019). Using as inputs not only isotopic measurements but also their uncertainties, the SIAR model can be used to

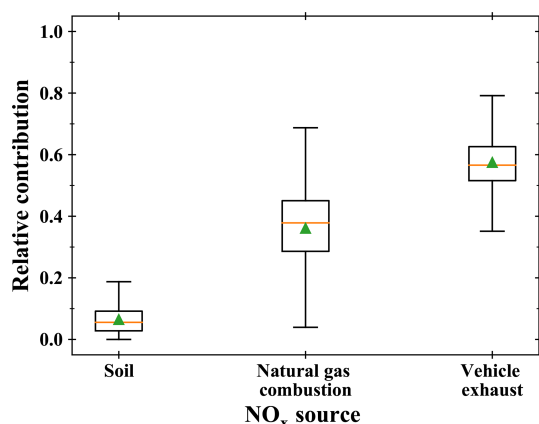
calculate potential NO<sub>x</sub> source solutions as probability distributions. A recent emission inventory of NO<sub>x</sub> in the Grenoble urban area estimated that, in 2016, 52 % of emitted NO<sub>x</sub> could be attributed to transport, 26 % to industries, 20 % to the residential/tertiary sectors, and 2 % to agriculture (Topin et al., 2019). Looking at the type of energy consumed by each sector, we estimate that at this time of the year, our sampling site was mostly influenced by fossil-fuel combustion NO<sub>x</sub> sources, mainly gasoline/diesel and natural gas, and by biogenic NO<sub>x</sub> sources (soils emissions). As shown by previous studies,  $\delta^{15}\text{N}$  of NO<sub>x</sub> emitted by vehicle exhausts depends on the fuel type, the reduction emission technology, and the vehicle run time with values ranging from −21‰ to −2‰ (Walters et al., 2015). As 90 % of traffic-related NO<sub>x</sub> is emitted by diesel-powered engines in the Grenoble urban area (Atmo-Auvergne-Rhône-Alpes, 2018), we use a value of  $(-4.7 \pm 1.7)\text{‰}$ , representative of the U.S. vehicle fleet (Miller et al., 2017) for which about 80 % of its traffic-NO<sub>x</sub> emissions originate from diesel vehicles (Dallmann et al., 2013). For  $\delta^{15}\text{N}$  of NO<sub>x</sub> emitted by natural gas combustion, we use a value of  $(-16.5 \pm 1.7)\text{‰}$  which is the average isotopic signature of natural-gas-burning power plants and residential furnace exhausts (Walters et al., 2015). Despite the large range of  $\delta^{15}\text{N}$  values for biogenic NO<sub>x</sub> (from −59.8 % to −19.9 %) (Li and Wang, 2008; Yu and Elliott, 2017; Walters et al., 2015), these values are still very distinct from  $\delta^{15}\text{N}$  of fossil-fuel combustion NO<sub>x</sub>, making possible to roughly estimate the relative contributions of different NO<sub>x</sub> sources at our sampling site. We use a soil-NO<sub>x</sub>  $\delta^{15}\text{N}$  value of  $(-33.8 \pm 12.2)\text{‰}$  (Zong et al., 2017), which is the average of values taken from several studies on NO<sub>x</sub> emitted by natural and fertilized soil (Felix and Elliott, 2014; Li and Wang, 2008). Over our sampling period, the SIAR model results indicate traffic as the dominant NO<sub>x</sub> emission source with a mean relative contribution of  $(57 \pm 8)\%$  (see Fig. 5). Natural gas combustion is found to be the second main NO<sub>x</sub> emission source  $((36 \pm 12)\%)$  before soil emissions which account for only  $(7 \pm 5)\%$ . The limited nature of our measurement dataset (only 1 d of sampling) prevents us from drawing any general and robust conclusions on the relative contributions of NO<sub>x</sub> emissions at our site. Nonetheless, we note that the SIAR overall source apportionment is in close agreement with the Grenoble urban area emission inventory concerning traffic emissions (52 % in 2016), lending some support to the idea that  $\delta^{15}\text{N}$  of NO<sub>2</sub> is a reliable tracer of NO<sub>x</sub> emission sources after correction for LCIE and EIE.

## 5 Conclusion

The primary goal of this preliminary work was to address an efficient and portable sampling system for atmospheric NO<sub>2</sub> fitting with accurate isotopic analysis of double nitrogen and triple oxygen isotopes. First simultaneous measurements of the multi-isotopic composition ( $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ ,

**Table 3.** Summary of measured  $f_{\text{NO}_2}$ , calculated  $A^*$  values using Eq. (9) for daytime and Eq. (11) for nighttime, calculated isotopic fractionation between  $\text{NO}_2$  and  $\text{NO}_x$  ( $\Delta(\text{NO}_2 - \text{NO}_x)$ ) using Eq. (8) for daytime and Eq. (10) for nighttime, and  $\delta^{15}\text{N}(\text{NO}_x)$  estimated with  $\Delta(\text{NO}_2 - \text{NO}_x)$  and measured  $\delta^{15}\text{N}(\text{NO}_2)$ .

Sampling date & time (start–end)	$f_{\text{NO}_2}$	$A^*$	$\Delta(\text{NO}_2 - \text{NO}_x)$ (‰)	$\delta^{15}\text{N}(\text{NO}_x)$ (‰)
14 May 2019 21:00–00:00	1.00	1.70	0.00	−11.70
15 May 2019 06:00–09:00	0.87	0.27	2.72	−7.62
15 May 2019 09:00–12:00	0.85	1.46	0.85	−10.95
15 May 2019 12:00–15:00	0.81	2.61	0.01	−11.81
15 May 2019 15:00–18:00	0.95	3.38	−0.13	−10.87
15 May 2019 18:00–21:00	1.00	3.04	0.00	−11.22
16 May 2019 00:00–05:00	1.00	0.42	0.00	−11.10



**Figure 5.** Potential  $\text{NO}_x$  emission source partitioning using the SIAR model based on estimated  $\delta^{15}\text{N}(\text{NO}_x)$ . Reference values for each source were taken from Miller et al. (2017), Walters et al. (2015), and Zong et al. (2017).

and  $\Delta^{17}\text{O}$ ) of atmospheric  $\text{NO}_2$  are reported here, notably at relatively high temporal resolution (3 h). Over the course of more than 1 d in the Grenoble urban/suburban environment,  $\Delta^{17}\text{O}(\text{NO}_2)$  is found to vary diurnally with a maximum value of  $(39.2 \pm 0.3)\text{‰}$  during the day and a minimum value of  $(20.5 \pm 0.3)\text{‰}$  at night. At photo-stationary state, high  $\Delta^{17}\text{O}(\text{NO}_2)$  values result from the ozone predominance in NO oxidation pathways, whereas lower values reflect the influence of peroxy radicals. We estimate from our  $\Delta^{17}\text{O}(\text{NO}_2)$  measurements that 86 % of  $\text{NO}_2$  produced by day originates from the oxidation of NO by  $\text{O}_3$ . Moreover, a mean daytime peroxy radical mixing ratio of  $(13.8 \pm 11.2)\text{pmol mol}^{-1}$  is derived from the oxygen isotopic measurements. At night,  $\text{NO}_x$  photochemistry shuts down; hence,  $\Delta^{17}\text{O}(\text{NO}_2)$  decreases under the growing influence of the isotopic footprint from NO emitted by night. The  $\Delta^{17}\text{O}(\text{NO}_2)$  measurement towards the end of the night is found to be quantitatively consistent with typical values of  $\Delta^{17}\text{O}(\text{O}_3)$ . The  $\delta^{15}\text{N}(\text{NO}_2)$  measurements show little variations, from  $-11.8\text{‰}$  to  $-4.9\text{‰}$ , with mostly negligible N

isotope fractionations between NO and  $\text{NO}_2$  due to the high  $\text{NO}_2 / \text{NO}_x$  ratios. After correction of nitrogen isotopic fractionations, we use a Bayesian isotope mixing model to estimate the relative contributions of the dominant  $\text{NO}_x$  emissions sources. The results indicate the predominance of traffic  $\text{NO}_x$  emissions in this area at  $(57 \pm 8)\%$ , which is before natural gas combustion and soil emission.

Despite the limited nature of our measurement dataset, our results shed light on the sensitivity of  $\text{NO}_2$  isotopic signature to the atmospheric chemical regimes and emissions of the local environment. This isotopic approach can be applied to various environments in order to probe further the oxidative chemistry and help to constrain the  $\text{NO}_x$  fate in a more quantitative way. In the future, the interpretation of the isotopic data should be extended with the use of a photochemical box model including isotopic anomaly transfers and local emissions in order to solve persistent issues of atmospheric oxidation mechanisms. Moreover, samplings and multi-isotopic analysis of atmospheric nitrate performed in parallel to those of  $\text{NO}_2$  would certainly be of interest for the study of the full reactive nitrogen cycle.

## Appendix A: Isotopic standards and calibration

This method of analysis induces isotope fractionations during  $\text{NO}_2^-/\text{N}_2\text{O}$  conversion and ionization in the spectrometer, as well as isotope exchanges between  $\text{NO}_2^-$  and its medium. Indeed, while isotope exchanges between nitrite and its matrix are minimized due to the basic pH, the chemistry required to convert nitrite to  $\text{N}_2\text{O}$  involves a step in an acidic medium that promotes an exchange of oxygen isotopes (Casciotti et al., 2007). In order to eliminate the effects of these isotope splits, the system is calibrated using standards of known isotopic composition, which are subjected to the same treatment as the samples. This is called the identical treatment principle (Brand, 1996). By subjecting compounds of known isotopic composition to the same treatment as the samples, the isotope fractionation induced by the analytical protocol can be estimated, and the sample values can be corrected. Standards are first dissolved in a basic aqueous medium ( $\text{pH} = 12$ ), and then, from this stock solution, five series of each standard are prepared in several amount ranges, namely 40, 80, 100, 120, and 150 nmol, in order to estimate the effects of the amount of a material on its isotopic measurement. The matrix used for their preparation is the same as that of the samples, i.e. a mixture of KOH 2 M/guaiacol in deionized water. Correction factors are obtained by linear regression between the raw and the expected values of  $\delta^{15}\text{N}$ ,  $\delta^{18}\text{O}$ , and  $\delta^{17}\text{O}$  of the standards. Three international references of known  $\delta^{15}\text{N}$  and  $\delta^{18}\text{O}$  values are used for this work. These are nitrite salts, named RSIL-N7373, RSIL-N10219, and RSIL-N23 with respective  $\delta^{15}\text{N}/\delta^{18}\text{O}$  values of  $-79.6/4.2\text{‰}$ ,  $2.8/88.5\text{‰}$ , and  $3.7/11.4\text{‰}$ . Although the three standards cover a wide range of isotopic composition in  $\delta^{15}\text{N}$  and  $\delta^{18}\text{O}$ , they do not have an isotopic anomaly in  $^{17}\text{O}$ . As we are not aware of any available international reference nitrite standards with a known  $^{17}\text{O}$  anomaly, we are currently in the process of manufacturing our own standards. As this step is still under development and in order to be able to assess the accuracy of our  $^{17}\text{O}$  measurements of atmospheric  $\text{NO}_2$  samples, we have estimated the isotope fractionation that  $^{17}\text{O}$  undergoes during the analysis. RSIL-N7373 and RSIL-N23 standards have a  $\Delta^{17}\text{O} = 0\text{‰}$ , so we estimate their  $^{17}\text{O}$  composition such that  $\delta^{17}\text{O} = 0.52 \times \delta^{18}\text{O}$ . For standard RSIL-N10219, we measure a negative  $\Delta^{17}\text{O}$  around  $-7\text{‰}$ . We therefore apply the mass-independent relation such that  $\delta^{17}\text{O}_{\text{std}}(\text{RSIL-N10219}) = \Delta^{17}\text{O}_{\text{raw}}(\text{RSIL-N10219}) + 0.5 \times \delta^{18}\text{O}_{\text{std}}(\text{RSIL-N10219})$ .

The isotopic exchange of  $^{18}\text{O}$  is estimated at 11 % for standards at 100 nmol (Fig. A1), which is in line with Kobayashi et al. (2021), who have estimated the degree of O isotope exchange in the azide method between  $\text{H}_2\text{O}$  and  $\text{NO}_2^-$  to be  $(10.8 \pm 0.3)\%$ . The  $^{15}\text{N}$  calibration curve allows us to ensure a good fractionation rate during the analysis. Indeed, given the 1 : 1 association of the nitrogen atoms of nitrite and azide, the theoretical value of the calibration slope must be

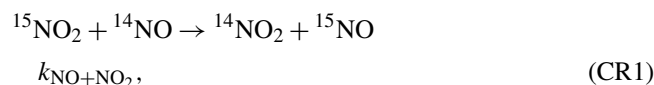
0.5. The slight deviation from our measured value can be attributed to a blank effect, estimated here at 2 % of the size of the standards (6 % for those at 40 nmol).

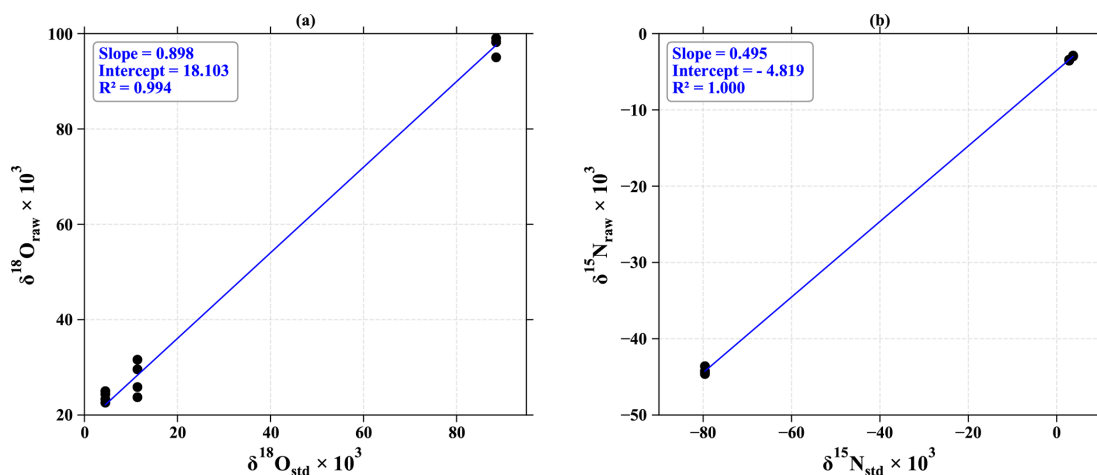
## Appendix B: Sample's isotopic stability

Oxygen isotopes in nitrites are very labile (Böhlke et al., 2007), but the basic pH of the eluent limits isotopic exchanges. To ensure isotopic integrity from denuder's extraction to analysis by IRMS, we followed the procedure by Walters et al. (2018) to quantify isotopic exchanges that might occur with the eluted matrix during storage. Thus, three solutions each containing 500 nmol of a  $\text{KNO}_2$  salt (RSIL-N7373, RSIL-N10219 and RSIL-N23) were prepared in the eluted matrix and kept frozen. We monitored the nitrite standard isotopic composition prepared in the eluted guaiacol matrix during 22 d. To do so, 100 nmol were collected from the individual solutions, analysed, and refrozen until the next analysis. The temporal evolution of the  $\delta^{17}\text{O}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$  differences between our measurements of RSIL standards (prepared in the KOH/guaiacol eluted matrix) and their certified reference values is plotted in Fig. B1. It represents the temporal drift of the isotopic signal with respect to reference values. If the deviation is constant, it means that the isotopic signal is not degraded with time, and its standard deviation is considered the uncertainty in our  $\delta^{17}\text{O}(\text{NO}_2)$  and  $\delta^{18}\text{O}(\text{NO}_2)$  measurements. As shown in Fig. B1, deviation of the three standards was stable over the 22 d experiment with an overall mean of  $(1.1 \pm 0.8)\text{‰}$ ,  $(2.3 \pm 1.8)\text{‰}$ , and  $(-0.1 \pm 0.3)\text{‰}$  for  $\delta^{17}\text{O}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$ , respectively. Note that RSIL-N10219 shows higher  $\delta^{17}\text{O}$  and  $\delta^{18}\text{O}$  residuals than the two other standards. The reason for this difference in behaviour is still not fully understood. As residuals remain steady over several weeks, we consider this method suitable for the oxygen analysis of  $\text{NO}_2$ , and the uncertainties applied to our isotopic measurements are reported as the propagation error of the mean measurement uncertainty and the mean uncertainty resulting from  $\text{NO}_2^-$  storage. In our study, average uncertainties on  $\delta^{17}\text{O}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$  are estimated to be  $\pm 0.8\text{‰}$ ,  $\pm 1.8\text{‰}$ , and  $\pm 0.3\text{‰}$ , respectively ( $1\sigma$  uncertainties).

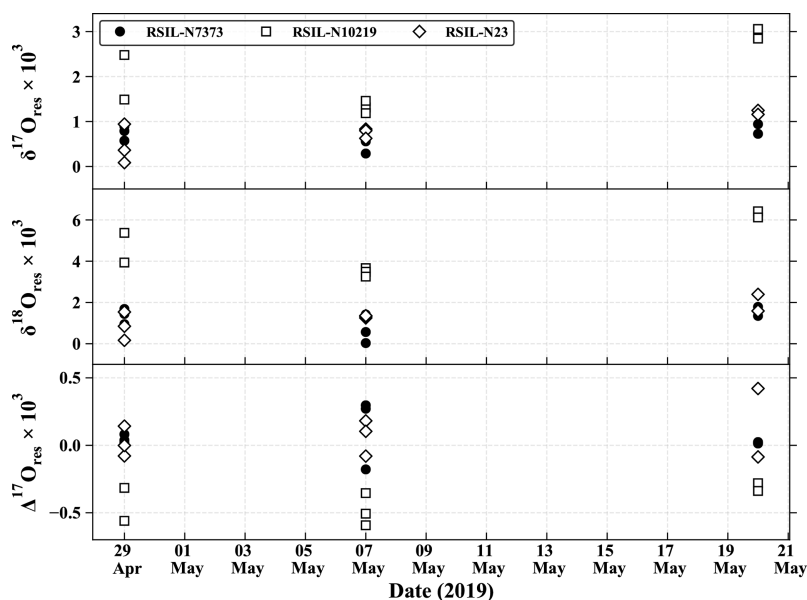
## Appendix C: Deriving the N isotopic fractionation from isotopic exchange and the extended Leighton cycle

We follow the same approach as Li et al. (2020) but take into account all the oxidation pathways of NO into  $\text{NO}_2$ , not only via  $\text{O}_3$ . The reactions considered in deriving the combined isotopic fractionation are the following:

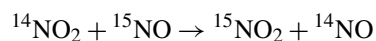




**Figure A1.** Calibration of (a)  $^{18}\text{O}$  and (b)  $^{15}\text{N}$  with nitrite standards at 100 nmol measured by the chemical azide method. The measured  $\delta^{18}\text{O}$  ( $\delta^{18}\text{O}_{\text{raw}}$ ) and  $\delta^{15}\text{N}$  ( $\delta^{15}\text{N}_{\text{raw}}$ ) values of  $\text{NO}_2^-$  standards are plotted against their certified reference  $\delta^{18}\text{O}$  ( $\delta^{18}\text{O}_{\text{std}}$ ) and  $\delta^{15}\text{N}$  ( $\delta^{15}\text{N}_{\text{std}}$ ) values.

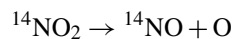


**Figure B1.** Temporal evolution of  $\delta^{17}\text{O}$ ,  $\delta^{18}\text{O}$ , and  $\Delta^{17}\text{O}$  differences between our measurements of RSIL standards (prepared in the KOH/guaiacol eluted matrix) and their certified reference values. Error bars derived from measurement uncertainties are approximately equivalent to the size of the markers.



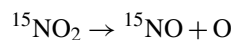
$$k_{\text{NO}+\text{NO}_2} \times \alpha_{\text{EIE}},$$

(CR2)



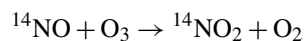
$$J_{\text{NO}_2},$$

(CR3)



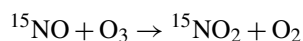
$$J_{\text{NO}_2} \times \alpha_{\text{PHIFE}},$$

(CR4)



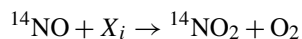
$$k_{\text{NO}+\text{O}_3},$$

(CR5)



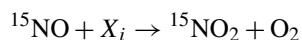
$$k_{\text{NO}+\text{O}_3} \times \alpha_{\text{KIE}(\text{NO}+\text{O}_3)},$$

(CR6)



$$k_{\text{NO}+X_i},$$

(CR7)



$$k_{\text{NO}+X_i} \times \alpha_{\text{KIE}(\text{NO}+X_i)},$$

(CR8)

with  $X_i = \text{RO}_2, \text{BrO}, \text{ClO}, \dots$

### C1 Daytime N fractionation

During the day, NO<sub>2</sub> photolysis is the overwhelmingly dominant NO<sub>2</sub> sink, and NO oxidation is the main NO<sub>2</sub> source (see Fig. C1). The assumption of steady state on <sup>15</sup>NO<sub>2</sub> for the extended Leighton cycle leads to

$$k_{\text{NO}+\text{NO}_2} [^{15}\text{NO}_2] [^{14}\text{NO}] + J_{\text{NO}_2} \alpha_{\text{PHIFE}} [^{15}\text{NO}_2] = \sum (k_{\text{NO}+X_i} \alpha_{\text{KIE}(\text{NO}+X_i)} [^{15}\text{NO}] [X_i]) + k_{\text{NO}+\text{NO}_2} \alpha_{\text{EIE}} [^{14}\text{NO}_2] [^{15}\text{NO}], \quad (\text{C1})$$

where  $k_{\text{NO}+\text{NO}_2}$  is the rate constant for the nitrogen isotopic exchange between NO and NO<sub>2</sub>,  $J_{\text{NO}_2}$  is the NO<sub>2</sub> photolysis rate with  $\alpha_{\text{PHIFE}}$  being its isotopic fractionation factor,  $\sum k_{\text{NO}+X_i} [X_i]$  the sum of all the NO oxidation pathways to NO<sub>2</sub>,  $X_i$  the NO oxidant (i.e. O<sub>3</sub>, RO<sub>2</sub>, BrO, ClO ...), and  $k_{\text{NO}+X_i}$  the rate constant for the reaction of NO +  $X_i$  with  $\alpha_{\text{KIE}(\text{NO}+X_i)}$  being its isotopic fractionation factor. Equation (C1) can be rearranged to give

$$\frac{[^{15}\text{NO}_2]}{[^{15}\text{NO}]} = \frac{\sum (k_{\text{NO}+X_i} \alpha_{\text{KIE}(\text{NO}+X_i)} [X_i]) + k_{\text{NO}+\text{NO}_2} \alpha_{\text{EIE}} [^{14}\text{NO}_2]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}] + J_{\text{NO}_2} \alpha_{\text{PHIFE}}}. \quad (\text{C2})$$

Meanwhile, <sup>14</sup>NO<sub>2</sub> in steady state leads to

$$\frac{[^{14}\text{NO}_2]}{[^{14}\text{NO}]} = \frac{\sum k_{\text{NO}+X_i} [X_i]}{J_{\text{NO}_2}}. \quad (\text{C3})$$

We define  $A_{\text{day}}^*$  as the ratio of the <sup>14</sup>NO<sub>2</sub> lifetime with respect to isotopic exchange with <sup>14</sup>NO ( $\tau_{\text{exchange-NO}_2}$ ) over the daytime <sup>14</sup>NO<sub>2</sub> chemical lifetime ( $\tau_{\text{chem-NO}_2}$ ) (Li et al., 2020):

$$A_{\text{day}}^* = \frac{\tau_{\text{exchange-NO}_2}}{\tau_{\text{chem-NO}_2}} = \frac{J_{\text{NO}_2}}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}]}. \quad (\text{C4})$$

Using Eq. (C3), Eq. (C4) becomes

$$A_{\text{day}}^* = \frac{\sum k_{\text{NO}+X_i} [X_i]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}_2]}. \quad (\text{C5})$$

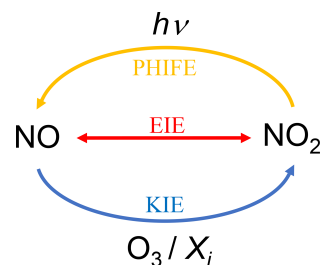
We also define  $T_{\text{NO}+X_i}$  as the relative importance of the oxidation pathway of NO into NO<sub>2</sub> via the oxidant  $X_i$ :

$$T_{\text{NO}+X_i} = \frac{k_{\text{NO}+X_i} [X_i]}{\sum k_{\text{NO}+X_i} [X_i]}, \quad (\text{C6})$$

with necessarily  $\sum T_{\text{NO}+X_i} = 1$ .

Using the definitions (C5) and (C6), Eq. (C2) becomes

$$\frac{[^{15}\text{NO}_2]}{[^{15}\text{NO}]} = \frac{A_{\text{day}}^* k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}_2] \alpha_{\text{KIE}} + k_{\text{NO}+\text{NO}_2} \alpha_{\text{EIE}} [^{14}\text{NO}_2]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}] + A_{\text{day}}^* k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}] \alpha_{\text{PHIFE}}}, \quad (\text{C7})$$



**Figure C1.** Sketch of the nitrogen fractionation processes between NO and NO<sub>2</sub>. PHIFE represents the photochemical isotope fractionation effect, KIE represents the kinetic isotope effect, and EIE represents the equilibrium isotope effect.

with  $\alpha_{\text{KIE}} = \sum T_{\text{NO}+X_i} + \alpha_{\text{KIE}(\text{NO}+X_i)}$ . Using  $R(^{15}\text{N}/^{14}\text{N}, \text{NO}) = R_{\text{NO}}(^{15}\text{N})/R_{\text{NO}}(^{14}\text{N}) = {}^{15}R_{\text{NO}}$  (with  $\delta^{15}\text{N}(\text{NO}) = {}^{15}R_{\text{NO}}/{}^{15}R_{\text{standard}} - 1$ ) and  $R(^{15}\text{N}/^{14}\text{N}, \text{NO}_2) = R_{\text{NO}_2}(^{15}\text{N})/R_{\text{NO}_2}(^{14}\text{N}) = {}^{15}R_{\text{NO}_2}$  (with  $\delta^{15}\text{N}(\text{NO}_2) = {}^{15}R_{\text{NO}_2}/{}^{15}R_{\text{standard}} - 1$ ), Eq. (C7) becomes

$$\frac{{}^{15}R_{\text{NO}_2}}{{}^{15}R_{\text{NO}}} = \frac{[^{15}\text{NO}_2][^{14}\text{NO}]}{[^{15}\text{NO}][^{14}\text{NO}_2]} = \frac{A_{\text{day}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}}{1 + A_{\text{day}}^* \alpha_{\text{PHIFE}}}, \quad (\text{C8})$$

$$\frac{{}^{15}R_{\text{NO}}}{{}^{15}R_{\text{NO}_2}} - 1 = \frac{A_{\text{day}}^* (\alpha_{\text{PHIFE}} - \alpha_{\text{KIE}}) - (\alpha_{\text{EIE}} - 1)}{A_{\text{day}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}}. \quad (\text{C9})$$

As a result, the daytime isotopic shift of NO<sub>2</sub> relative to NO, defined as  $\Delta(\text{NO}_2 - \text{NO}) = \delta^{15}\text{N}(\text{NO}_2) - \delta^{15}\text{N}(\text{NO})$ , is given by

$$\Delta_{\text{day}}(\text{NO}_2 - \text{NO}) = \frac{A_{\text{day}}^* (\alpha_{\text{KIE}} - \alpha_{\text{PHIFE}}) + (\alpha_{\text{EIE}} - 1)}{A_{\text{day}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}} \times (1 + \delta^{15}\text{N}(\text{NO}_2)). \quad (\text{C10})$$

Using the isotopic balance  $\delta^{15}\text{N}(\text{NO}_x) = f_{\text{NO}_2} \delta^{15}\text{N}(\text{NO}_2) + (1 - f_{\text{NO}_2}) \delta^{15}\text{N}(\text{NO})$  with  $f_{\text{NO}_2} = [\text{NO}_2]/[\text{NO}_x]$  (Li et al., 2020), the isotopic shift of NO<sub>2</sub> relative to NO<sub>x</sub>, defined as  $\Delta(\text{NO}_2 - \text{NO}_x) = \delta^{15}\text{N}(\text{NO}_2) - \delta^{15}\text{N}(\text{NO}_x)$ , can be expressed by

$$\Delta_{\text{day}}(\text{NO}_2 - \text{NO}_x) = \frac{A_{\text{day}}^* (\alpha_{\text{KIE}} - \alpha_{\text{PHIFE}}) + (\alpha_{\text{EIE}} - 1)}{A_{\text{day}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}} \times (1 + \delta^{15}\text{N}(\text{NO}_2)) (1 - f_{\text{NO}_2}). \quad (\text{C11})$$

Since fractionation factors are close to unity and  $1 + \delta^{15}\text{N}(\text{NO}_2) \approx 1$ , Eq. (C11) can be further simplified by keeping only the dominant terms (Li et al., 2020):

$$\Delta_{\text{day}}(\text{NO}_2 - \text{NO}_x) \approx \frac{\alpha_{\text{LCIE}}^* A_{\text{day}}^* + (\alpha_{\text{EIE}} - 1)}{A_{\text{day}}^* + 1} \times (1 - f_{\text{NO}_2}), \quad (\text{C12})$$



with  $\alpha_{\text{LCIE}}^* = \alpha_{\text{KIE}} - \alpha_{\text{PHIFE}}$ .

Considering the localization of our sampling site (urban mid-latitude area), only  $\text{NO} + \text{RO}_2$  and  $\text{NO} + \text{O}_3$  are thought to be significant as  $\text{NO}_2$  formation pathways; hence,  $\alpha_{\text{LCIE}}^*$  becomes

$$\alpha_{\text{LCIE}}^* = T_{\text{NO}+\text{O}_3} \times \alpha_{\text{KIE}(\text{NO}+\text{O}_3)} + T_{\text{NO}+\text{RO}_2} \times \alpha_{\text{KIE}(\text{NO}+\text{RO}_2)} - \alpha_{\text{PHIFE}}, \quad (\text{C13})$$

and Eq. (C5) becomes

$$A_{\text{day}}^* = \frac{k_{\text{NO}+\text{O}_3} [\text{O}_3] + k_{\text{NO}+\text{RO}_2} [\text{RO}_2]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}_2]}. \quad (\text{C14})$$

Equations (4) and (5) from Sect. 4.1.2 can be combined to give

$$k_{\text{NO}+\text{O}_3} [\text{O}_3] + k_{\text{NO}+\text{RO}_2} [\text{RO}_2] = \frac{\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)}{\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)} k_{\text{NO}+\text{O}_3} [\text{O}_3]. \quad (\text{C15})$$

Using Eq. (C15), Eq. (C14) becomes

$$A_{\text{day}}^* = \frac{k_{\text{NO}+\text{O}_3} [\text{O}_3]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}_2]} \left( \frac{\Delta^{17}\text{O}_{\text{NO}+\text{O}_3}(\text{NO}_2)}{\Delta^{17}\text{O}_{\text{day}}(\text{NO}_2)} \right). \quad (\text{C16})$$

We consider several particular cases. The first case is when  $\alpha_{\text{KIE}(\text{NO}+\text{RO}_2)} \approx \alpha_{\text{KIE}(\text{NO}+\text{O}_3)}$ . Previous studies found that the  $\text{NO} + \text{O}_3$  reaction falls within the family of “normal kinetic isotope fractionation” with the  $\text{NO}_2$  produced being depleted in  $^{15}\text{N}$  (Walters and Michalski, 2016) compared to residual reactant  $\text{NO}$ . To our knowledge, no such experiment has been carried out for the  $\text{NO} + \text{RO}_2$  reaction. Nonetheless, considering the very close, and both very low, activation energies for the reaction  $\text{NO} + \text{O}_3$  and  $\text{NO} + \text{RO}_2$ , it is quite likely that the fractionation factors of these two reactions are similar. It follows that we obtain the same expression for  $\alpha_{\text{LCIE}}^*$  as for the  $\alpha_{\text{LCIE}}$  given in Li et al. (2020):

$$\alpha_{\text{LCIE}}^* = \alpha_{\text{KIE}(\text{NO}+\text{O}_3)} - \alpha_{\text{PHIFE}}. \quad (\text{C17})$$

And Eq. (C12) becomes

$$\Delta_{\text{day}}(\text{NO}_2 - \text{NO}_x) = \frac{\alpha_{\text{LCIE}}^* A_{\text{day}}^* + (\alpha_{\text{EIE}} - 1)}{A_{\text{day}}^* + 1} \times (1 - f_{\text{NO}_2}). \quad (\text{C18})$$

Equation (C18) with  $A_{\text{day}}^*$  given by Eq. (C16) is the expression that we use to analyse our daytime nitrogen isotopic measurements. Another particular case considered by Li et al. (2020) is  $k_{\text{NO}+\text{O}_3} [\text{O}_3] \gg k_{\text{NO}+\text{RO}_2} [\text{RO}_2]$ ; in that case,  $\alpha_{\text{LCIE}}^*$  is still given by Eq. (C12) but  $A^*$  is simplified:

$$A_{\text{day}}^* = \frac{k_{\text{NO}+\text{O}_3} [\text{O}_3]}{k_{\text{NO}+\text{NO}_2} [\text{NO}_2]}. \quad (\text{C19})$$

Equation (C16) with  $A_{\text{day}}^*$  given by Eq. (C19) and  $\alpha_{\text{LCIE}}^*$  given by Eq. (C17) is the same expression as Eq. (8) in Li et al. (2020).

## C2 Nighttime N fractionation

An expression similar to Eq. (C1) can be derived for nighttime conditions, when  $\text{NO}_2$  photolysis is null and hence when there is no recycling between  $\text{NO}$  and  $\text{NO}_2$ . In addition, the conversion of  $\text{NO}$  into  $\text{NO}_2$  occurs only via reaction with  $\text{O}_3$  because the mixing ratios of other  $\text{NO}$  oxidants are usually negligible at night. The main source of  $\text{NO}_x$  at night is the  $\text{NO}$  emissions. The assumption of steady state on short timescales can only hold for  $^{14}\text{NO}$  and  $^{15}\text{NO}$ , not  $\text{NO}_2$ , leading to an equation equivalent to Eq. (C1):

$$k_{\text{NO}+\text{NO}_2} [^{15}\text{NO}_2] [^{14}\text{NO}] + E(^{15}\text{NO}) = k_{\text{NO}+\text{O}_3} \alpha_{\text{KIE}} [^{15}\text{NO}] [\text{O}_3] + k_{\text{NO}+\text{NO}_2} \alpha_{\text{EIE}} [^{14}\text{NO}_2] [^{15}\text{NO}], \quad (\text{C20})$$

with  $E(^{15}\text{NO})$  being the  $^{15}\text{NO}$  emission flux and  $\alpha_{\text{KIE}}$  is the fractionation factor of  $\text{NO} + \text{O}_3$ . Equation (C20) can be rearranged to give

$$\frac{[^{15}\text{NO}_2]}{[^{15}\text{NO}]} = \frac{k_{\text{NO}+\text{O}_3} \alpha_{\text{KIE}} [\text{O}_3] + k_{\text{NO}+\text{NO}_2} \alpha_{\text{EIE}} [^{14}\text{NO}_2]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}] + E(^{15}\text{NO}) / [^{15}\text{NO}_2]}. \quad (\text{C21})$$

Meanwhile, the steady state on  $^{14}\text{NO}$  gives (nitrogen isotopic exchanges are neglected as  $\text{NO}$  emissions are largely dominated by  $^{14}\text{NO}$ )

$$E(^{14}\text{NO}) = k_{\text{NO}+\text{O}_3} [^{14}\text{NO}] [\text{O}_3], \quad (\text{C22})$$

with  $E(^{14}\text{NO})$  being the  $^{14}\text{NO}$  emission flux. For nighttime, we define  $A_{\text{night}}^*$  as the ratio of the  $^{14}\text{NO}$  lifetime with respect to isotopic exchange with  $^{14}\text{NO}_2$  ( $\tau_{\text{exchange-NO}}$ ) over the nighttime  $^{14}\text{NO}$  chemical lifetime ( $\tau_{\text{chem-NO}}$ ):

$$A_{\text{night}}^* = \frac{\tau_{\text{exchange-NO}}}{\tau_{\text{chem-NO}}} = \frac{k_{\text{NO}+\text{O}_3} [\text{O}_3]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}_2]}. \quad (\text{C23})$$

Using Eq. (C22), Eq. (C23) gives

$$A_{\text{night}}^* = \frac{E(^{14}\text{NO})}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}] [^{14}\text{NO}_2]}. \quad (\text{C24})$$

$\text{NO}_x$  is overwhelmingly emitted in the form of  $\text{NO}$ . The isotopic signature of  $\text{NO}$  emissions can be characterized with  $^{15}R_{\text{NOemis}}$  and  $\delta^{15}\text{N}(\text{NOemis}) = ^{15}R_{\text{NOemis}} / ^{15}R_{\text{standard}} - 1$ . Using the definition of  $^{15}R_{\text{NOemis}}$ , Eqs. (C23), and (C24), Eq. (C21) becomes

$$\frac{[^{15}\text{NO}_2]}{[^{15}\text{NO}]} = \frac{A_{\text{night}}^* k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}_2] \alpha_{\text{KIE}} + k_{\text{NO}+\text{NO}_2} \alpha_{\text{EIE}} [^{14}\text{NO}_2]}{k_{\text{NO}+\text{NO}_2} [^{14}\text{NO}] + (^{15}R_{\text{NOemis}} E(^{14}\text{NO})) / [^{15}\text{NO}_2]}. \quad (\text{C25})$$

And then, using Eq. (C24), Eq. (C25) becomes

$$\frac{{}^{15}R_{\text{NO}_2}}{{}^{15}R_{\text{NO}}} = \frac{[{}^{15}\text{NO}_2][{}^{14}\text{NO}]}{[{}^{15}\text{NO}][{}^{14}\text{NO}_2]} = \frac{A_{\text{night}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}}{1 + A_{\text{night}}^* ({}^{15}R_{\text{NO}_{\text{emis}}}/{}^{15}R_{\text{NO}_2})}, \quad (\text{C26})$$

$$\frac{{}^{15}R_{\text{NO}}}{{}^{15}R_{\text{NO}_2}} - 1 = \frac{A_{\text{night}}^* (({}^{15}R_{\text{NO}_{\text{emis}}}/{}^{15}R_{\text{NO}_2}) - \alpha_{\text{KIE}}) - (\alpha_{\text{EIE}} - 1)}{A_{\text{night}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}}. \quad (\text{C27})$$

Following the approach used in the derivation of daytime isotopic shift, the nighttime isotopic shift of  $\text{NO}_2$  relative to NO is given by

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}) = \frac{A_{\text{night}}^* (\alpha_{\text{KIE}} - ({}^{15}R_{\text{NO}_{\text{emis}}}/{}^{15}R_{\text{NO}_2})) + (\alpha_{\text{EIE}} - 1)}{A_{\text{night}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}} \times (1 + \delta^{15}\text{N}(\text{NO}_2)). \quad (\text{C28})$$

Using the isotopic balance  $\delta^{15}\text{N}(\text{NO}_x) = f_{\text{NO}_2} \delta^{15}\text{N}(\text{NO}_2) + (1 - f_{\text{NO}_2}) \delta^{15}\text{N}(\text{NO})$ , the nighttime isotopic shift of  $\text{NO}_2$  relative to  $\text{NO}_x$ , can be expressed by

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x) = \frac{A_{\text{night}}^* \left( \alpha_{\text{KIE}} - \left( \frac{1 + \delta^{15}\text{N}(\text{NO}_{\text{emis}})}{1 + \delta^{15}\text{N}(\text{NO}_2)} \right) \right) + (\alpha_{\text{EIE}} - 1)}{A_{\text{night}}^* \alpha_{\text{KIE}} + \alpha_{\text{EIE}}} \times (1 + \delta^{15}\text{N}(\text{NO}_2)) (1 - f_{\text{NO}_2}), \quad (\text{C29})$$

where  $\delta^{15}\text{N}(\text{NO}_{\text{emis}})$  is the nitrogen isotopic composition of NO emissions.

Keeping the dominant terms, Eq. (C29) can be further simplified following the daytime derivation:

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x) \approx \frac{A_{\text{night}}^* \left( \alpha_{\text{KIE}} - \left( \frac{1 + \delta^{15}\text{N}(\text{NO}_{\text{emis}})}{1 + \delta^{15}\text{N}(\text{NO}_2)} \right) \right) + (\alpha_{\text{EIE}} - 1)}{A_{\text{night}}^* + 1} \times (1 - f_{\text{NO}_2}). \quad (\text{C30})$$

We consider two particular cases. When  $A_{\text{night}}^* \ll 1$  ( $k_{\text{NO}+\text{NO}_2}[{}^{14}\text{NO}_2] \gg k_{\text{NO}+\text{O}_3}[\text{O}_3]$ ), i.e. isotopic exchange much faster than NO oxidation, Eq. (C28) becomes

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}) = \frac{(\alpha_{\text{EIE}} - 1)}{\alpha_{\text{EIE}}} (1 + \delta^{15}\text{N}(\text{NO}_2)). \quad (\text{C31})$$

Keeping the dominant terms, Eq. (C31) can be simplified:

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}) \approx (\alpha_{\text{EIE}} - 1). \quad (\text{C32})$$

As expected, the nighttime isotopic shift of  $\text{NO}_2$  relative to NO depends only on the isotopic exchange fractionation in

that case. In the same way, Eq. (C29) becomes

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x) = \frac{(\alpha_{\text{EIE}} - 1)}{\alpha_{\text{EIE}}} \times (1 + \delta^{15}\text{N}(\text{NO}_2)) (1 - f_{\text{NO}_2}). \quad (\text{C33})$$

Keeping the dominant terms, Eq. (C33) can be simplified:

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}_x) \approx \frac{(\alpha_{\text{EIE}} - 1)}{\alpha_{\text{EIE}}} (1 - f_{\text{NO}_2}). \quad (\text{C34})$$

When  $A_{\text{night}}^* \gg 1$  ( $k_{\text{NO}+\text{NO}_2}[{}^{14}\text{NO}_2] \ll k_{\text{NO}+\text{O}_3}[\text{O}_3]$ ), i.e. NO oxidation much faster than isotopic exchange, Eq. (C28) yields

$$\Delta_{\text{night}}(\text{NO}_2 - \text{NO}) = 1 + \delta^{15}\text{N}(\text{NO}_2) - \left( \frac{1 + \delta^{15}\text{N}(\text{NO}_{\text{emis}})}{\alpha_{\text{KIE}}} \right), \quad (\text{C35})$$

leading to

$$1 + \delta^{15}\text{N}(\text{NO}) = \left( \frac{1 + \delta^{15}\text{N}(\text{NO}_{\text{emis}})}{\alpha_{\text{KIE}}} \right) \quad (\text{C36})$$

$${}^{15}R_{\text{NO}} = \frac{{}^{15}R_{\text{NO}_{\text{emis}}}}{\alpha_{\text{KIE}}}. \quad (\text{C37})$$

As expected, the nighttime isotopic shift of NO relative to NO emissions depends only on the isotopic fractionation factor of the  $\text{NO} + \text{O}_3$  reaction in that case. It is possible to estimate the nighttime isotopic shift of  $\text{NO}_2$  relative to NO or  $\text{NO}_x$  on long timescales by assuming crudely that  ${}^{14}\text{NO}_2$  is in steady state:

$$k_{\text{loss-NO}_2}[{}^{14}\text{NO}_2] = k_{\text{NO}+\text{O}_3}[{}^{14}\text{NO}][\text{O}_3], \quad (\text{C38})$$

with  $k_{\text{loss-NO}_2}$  representing the equivalent of a first-order rate constant. If the  ${}^{14}\text{NO}_2$  loss is a second-order reaction such as  $\text{NO}_2 + \text{O}_3$  loss, then  $k_{\text{loss-NO}_2} = k_{\text{NO}_2+\text{O}_3}[\text{O}_3]$ . In the same way, assuming that the  $\text{NO}_2$  oxidation into nitrate via  $\text{O}_3$  is not fractionating,  ${}^{15}\text{NO}_2$  in steady state gives

$$k_{\text{loss-NO}_2}[{}^{15}\text{NO}_2] = k_{\text{NO}+\text{O}_3} \alpha_{\text{KIE}} [{}^{15}\text{NO}][\text{O}_3]. \quad (\text{C39})$$

Using Eq. (C38), Eq. (C39) becomes

$${}^{15}R_{\text{NO}_2} = \alpha_{\text{KIE}} {}^{15}R_{\text{NO}}. \quad (\text{C40})$$

Using Eq. (C37), Eq. (C40) becomes

$${}^{15}R_{\text{NO}_2} = {}^{15}R_{\text{NO,emis}} \quad (\text{C41})$$

or

$$\delta^{15}\text{N}(\text{NO}_2) = \delta^{15}\text{N}(\text{NO}_{\text{emis}}). \quad (\text{C42})$$

Under those conditions (negligible isotopic exchange), a measurement of  $\delta^{15}\text{N}(\text{NO}_2)$  is a measurement of  $\delta^{15}\text{N}(\text{NO}_{\text{emis}})$ , preferably towards the end of the night in order for  ${}^{14}\text{NO}_2$  to move towards steady state.

In the text,  ${}^{14}\text{NO}_2$  and  ${}^{14}\text{NO}$  are referred to as  $\text{NO}_2$  and NO for convenience.

## Appendix D: Kinetic data used

Table D1. Rate constants used for calculations.

Reaction number	Reactions	Rate constants ( $\text{cm}^3 \text{mol}^{-1} \text{s}^{-1}$ )	References
(R3)	$\text{NO} + \text{O}_3 \rightarrow \text{NO}_2 + \text{O}_2$	$k_{\text{NO}+\text{O}_3} = 1.4 \times 10^{-12} \exp(-310(K)/T)$	Atkinson et al. (2004)
(R4)	$\text{NO} + \text{RO}_2 \rightarrow \text{NO}_2 + \text{RO}$	$k_{\text{NO}+\text{RO}_2} = 2.3 \times 10^{-12} \exp(360(K)/T)$	Atkinson et al. (2006)
(R6)	$\text{NO}_2 + \text{O}_3 \xrightarrow{M} \text{NO}_3 + \text{O}_2$	$k_{\text{NO}_2+\text{O}_3} = 1.4 \times 10^{-13} \exp(-2470(K)/T)$	Atkinson et al. (2004)
(CR1)	$^{15}\text{NO}_2 + ^{14}\text{NO} \rightarrow ^{14}\text{NO}_2 + ^{15}\text{NO}$	$k_{\text{NO}+\text{NO}_2} = 8.14 \times 10^{-14}$	Sharma et al. (1970)

**Code availability.** The Bayesian mixing model used for this study is available to download from the packages section of the Comprehensive R Archive Network site (CRAN; <https://CRAN.R-project.org/package=siar>, Parnell and Jackson, 2013).

**Data availability.** Data used for this study have been included in the Supplement.

**Supplement.** The supplement related to this article is available online at: <https://doi.org/10.5194/acp-21-10477-2021-supplement>.

**Author contributions.** Sampling and analysis protocol were developed by SA under the supervision of JS. NC and AB contributed with technical and knowledge support to SA for isotopic mass spectrometry and more general atmospheric measurements. SB and JS, supervisors of SA's PhD thesis, helped SA in interpreting the results and writing the article.

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