



The Role of Lightning in Atmospheric Chemistry

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Key Points:

- A review of literature published since 2007 on production of trace gases by lightning and their role in atmospheric chemistry is presented
- The best estimate of the global annual production of NO_x by lightning is $5 \pm 2 \text{ Tg N yr}^{-1}$, a narrower range than older estimates
- Direct production of O₃ in corona discharges and HO_x in corona and subvisible discharges should be considered in chemical models

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Abstract Lightning has important effects on atmospheric constituents in primarily the middle and upper troposphere. We summarize the effects of lightning on atmospheric chemistry that have been published since the last major review in 2007. The estimates of NO_x production by lightning (LNO_x) are based on five methods: theoretical/laboratory estimates applied to flash observations; aircraft measurements in storm anvils; cloud resolved modeling constrained by observations; satellite retrievals of NO₂ columns; and global and regional models constrained by aircraft or satellite data. The resulting best estimate of global LNO_x production is $5 \pm 2 \text{ Tg N yr}^{-1}$, which represents a small narrowing of the range from previous estimates. Model (observational) estimates are on average at the upper (lower) ends of this range. LNO_x production efficiency (PE) corresponding to this range is 150–350 mol fl⁻¹. An inverse relationship between flash rate and LNO_x PE has been ascertained from satellite and aircraft measurements. Direct O₃ and HO_x production in corona discharges and HO_x from subvisible flashes may be sufficiently large to warrant inclusion in models. LNO_x leads to photochemical O₃ production in the upper troposphere at rates of 5–10 ppbv d⁻¹ and contributes 1–2 ppbv to maximum 8-hr surface O₃ (greater amounts not uncommon over higher terrain). CH₄ lifetime decreases by ~5% per Tg N yr⁻¹ increase of LNO_x. Globally, no significant long-term trends in lightning flash rate have been observed, but regional trends have been detected. Future flash rate trends and LNO_x emissions remain uncertain.

Plain Language Summary We review the literature from the period 2007 to 2025 concerning the effects of lightning on atmospheric chemistry. The most well-known effect is the production of nitrogen oxides, which lead through chemical reactions in the presence of sunlight to production of ozone and to shortening the lifetime of methane, a potent greenhouse gas. We find that while the magnitude of nitrogen oxide production from lightning remains uncertain, the range of possible global production deduced through five techniques has been narrowed slightly since the last review of the subject in 2007. Other important findings are that nitrogen oxide production per flash is inversely related to the flash rate, and flash rates in global and regional models are best represented by schemes involving cloud and precipitation elements. Most of the ozone resulting from nitrogen oxide production by lightning occurs in the upper troposphere with only very small amounts delivered to the surface. No significant trends in lightning occurrence have been detected globally, but both positive and negative trends are present regionally.

1. Introduction

Lightning is a natural mechanism for the fixation of atmospheric nitrogen and a major source of nitrogen oxides (NO_x, i.e., NO (nitric oxide) and NO₂ (nitrogen dioxide)). The extreme temperatures (>15,000 K) in lightning channels dissociate molecular oxygen (O₂) and molecular nitrogen (N₂), forming NO when the channel cools (Zel'dovich & Raizer, 1967). This NO quickly reacts with ozone (O₃) to form NO₂, thereby an equilibrium state between NO and NO₂ is rapidly established. Lightning-induced NO_x (LNO_x) is one of the largest natural sources of NO_x in the atmosphere and the most prominent source of NO_x in the upper troposphere, particularly in the tropics and in the summertime midlatitudes. Globally, lightning is responsible for 10%–15% of all NO_x emissions. However, the regional contribution varies greatly as flash rates vary regionally and NO_x production per flash is highly variable. In the past few decades, the relative importance of natural sources of NO_x including lightning and soil has increased, especially over North America, Europe, and eastern China, where anthropogenic emissions have been decreasing. This is one of many reasons why LNO_x research is of high relevance and an update of the present knowledge since the publication by Schumann and Huntrieser in 2007 is needed. This review will provide that and includes an updated range of most likely values for LNO_x production after normalizing the literature

values to be in terms of Lightning Imaging Sensor (LIS) flash rates. The production of other trace gases by lightning is also considered in this review, as well as the effects of LNO_x on O_3 and CH_4 .

Other relevant reviews include the following. Rakov and Rachidi (2009) review a subset of publications over the 1998–2009 time period on lightning discharges, characteristics, and effects. Methods of detecting and geolocating flashes are reviewed in Nag et al. (2015). Verma et al. (2021) provide a review of 19 publications on lightning characteristics, lightning- NO_x production, and O_3 formation published between 2004 and 2019. Ng and Chou (2025) conducted a bibliometric review of lightning research. As part of that review, they used keyword analysis to break lightning research into seven clusters: (a) physical process of lightning, (b) lightning and climate interactions, (c) global and atmospheric lightning systems, (d) advanced lightning detection and prediction, (e) transient luminous events, (f) atmospheric convection and meteorology, and (g) environmental chemistry and lightning, the cluster that includes the impact of lightning- NO_x production on atmospheric composition.

Cloud-scale simulations of thunderstorms indicate that 55%–75% of LNO_x is detrained above 8 km (Pickering et al., 1998) where it leads to enhanced NO_y , O_3 , and OH, where the latter acts as an important oxidizing agent for many trace gases (Allen et al., 2010; Labrador et al., 2005; Liaskos et al., 2015). The produced O_3 enhances the longwave radiative absorption in the upper troposphere and therefore acts as an important greenhouse gas at those altitudes (Finney et al., 2018; Lacis et al., 1990). Conversely, the presence of elevated OH leads to a reduction in CH_4 lifetime and therefore a decrease in longwave radiative absorption (Finney et al., 2018; Fiore et al., 2006). The lifetime of NO_x in the upper troposphere is controlled by the chemical cycling of NO_x with reservoir species and is on average 10–20 days without the presence of deep convection (Prather & Jacob, 1997), but only 2–12 hr in the vicinity of convection (Nault et al., 2016, 2017). This chemical recycling provides a prominent LNO_x source downwind of thunderstorms, which causes the O_3 production efficiency to be 4–20 times higher in the upper troposphere than at the surface, resulting in a disproportionate impact of LNO_x on the tropospheric O_3 budget (Grewe et al., 2001; Pickering et al., 1990; Sauvage et al., 2007) and atmospheric composition in general. Furthermore, LNO_x impacts air quality and also nitrogen deposition (Allen et al., 2012; Kaynak et al., 2008). On average LNO_x adds 1–2 ppbv to the surface O_3 budget (Kang, Pickering, et al., 2019) over the U. S., although contributions as large as 18 ppbv have been observed during individual events (Murray, 2016). Tost (2017) and Luhar et al. (2024) have found substantial effects of lightning on the atmospheric nitrate burden.

The distribution of lightning is fairly well-known over much of the Earth due to remote sensing observations and an increase in the number and capability of ground-based lightning networks (see Section 2). However, the LNO_x production efficiency (PE, mol fl^{-1}) is a major source of uncertainty and discussed in detail later (Section 3). Thus, the global amount of NO_x due to lightning, which cannot be measured directly, is difficult to determine. LNO_x PE varies with flash type (cloud-to-ground (CG) versus intracloud (IC) flashes), flash length, multiplicity (strokes or groups per flash), and flash energy or peak current. Several different methods have been used to estimate the NO_x production per flash and the global LNO_x source. This paper focuses on the application and results of these methods:

- Theoretical considerations, along with laboratory results, have been applied to observed flashes to estimate LNO_x production per flash.
- Field studies have included aircraft transects through thunderstorm anvils (Figure 1), and the measured NO_x data has been combined with observed flash rates to estimate LNO_x PE.
- Cloud-resolving models with parameterized flashes, constrained by observed radar reflectivity, aircraft NO/NO_2 observations, and flash rates have been used to obtain a representation of LNO_x within entire storms. Combining model results with field studies has proven useful because anvil measurements are incomplete and likely biased as current research aircraft are unable to sample the most active portions of thunderstorms.
- UV/VIS spectrometers flying on satellites provide radiances from which NO_2 column amounts can be retrieved over thunderstorms and combined with lightning observations. Model estimates of the NO_x/NO_2 ratio can then be used to estimate LNO_x amounts in storms in which flashes are observed.
- Global LNO_x production rates can be obtained from global chemical transport models constrained by satellite and aircraft observations.

Only in recent years, with the advent of satellite observations of lightning flashes and improved coverage by ground-based lightning networks, has there been sufficient data to make estimates of trends in the occurrence of lightning. No long-term trend in global flash rates has been found so far (Füllekrug et al., 2022), however clear regional trends have been observed in the contiguous United States (U.S.; e.g., Koshak et al., 2015), in China

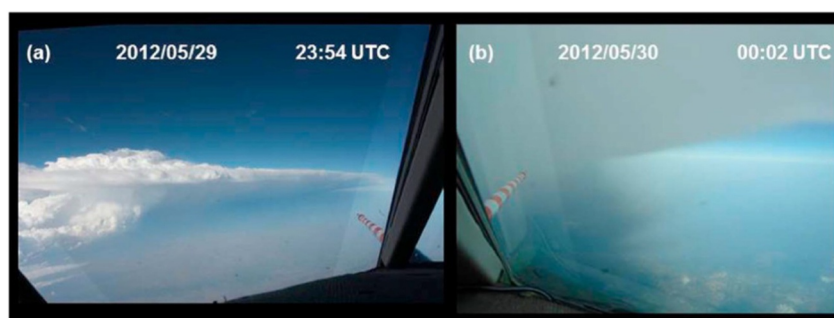


Figure 1. Falcon flight on 29–30 May 2012 during the Deep Convective Clouds and Chemistry (DC3) experiment: view from the cockpit (a) on 29 May at 23:54 UTC and (b) on 30 May at 00:02 UTC. The lower boundary of the anvil outflow from severe convection over Oklahoma was probed (Photo: R. Welser and A. Minikin, DLR; from Huntrieser et al., 2016b).

(H. Wang et al., 2021), over India (Prasanna et al., 2024), and in the Arctic (Holzworth et al., 2021). Though, it is yet unknown whether trends in total LNO_x production are in agreement with trends in lightning occurrence. Lightning characteristics may vary over time and between regions and therefore may have effects on the magnitude of LNO_x production. More data and time are needed to determine if local trends in flash rate are contributing to trends in LNO_x PE. More details on lightning trends are provided in Section 7.

In 2007, Schumann and Huntrieser compiled a comprehensive review of the literature concerning LNO_x production. They concluded that the global LNO_x production is $5 \pm 3 \text{ Tg N yr}^{-1}$. This global production rate is equivalent to $\sim 250 \text{ mol fl}^{-1}$ using the satellite-based flash climatology of Cecil et al. (2014), which indicates a global flash rate of 46 flashes s^{-1} . Therefore, the LNO_x source is larger than the global aircraft NO_x emissions estimated to be $\sim 1.4 \text{ Tg N yr}^{-1}$ (Lee et al., 2021; Teoh et al., 2024), which is the other major source of NO_x to the upper troposphere. In the 18 years since the Schumann and Huntrieser literature review, considerable research has been performed in attempts to narrow the LNO_x source uncertainty. This paper briefly summarizes the Schumann and Huntrieser (2007) findings as an introduction to the topic (Section 2) and thereafter reviews the present knowledge of LNO_x PE and the global LNO_x source rate based on the literature from the last 18 years (Section 3). Section 4 reviews the production of other trace gases (e.g., HO_x , O_3) in electrically active storms, where a considerable number of novel measurements are available and referenced (e.g., Brune et al., 2021). In Section 5 we review the magnitude of photochemical production of tropospheric O_3 downwind of specific thunderstorms and over broader scales. A summary of the effects of LNO_x on O_3 and CH_4 and ultimately on climate, as determined in global models, is provided in Section 6. Section 7 discusses observed trends in lightning flashes and model predictions for changes in future flash rates. In the last section, Section 8, a summary and conclusion on the current knowledge of LNO_x and other trace gas productivity by lightning and their effects on atmospheric chemistry is given.

2. Background Information

2.1. Types of Lightning Observations

For a detailed introduction to lightning and its observation types, we refer to Section 2.3 in Schumann and Huntrieser (2007). In this section, different types of lightning detection systems are summarized along with their use to estimate LNO_x PE.

Lightning is characterized by measurements of its position, type (mainly cloud-to-ground (CG) or intra-cloud (IC)), polarity, duration, peak current and energy. It is detected from the ground through VLF electromagnetic waves that are measurable thousands of km from the discharge. In addition, it can be detected on a regional scale by the brief VHF pulses ($\sim 1 \text{ ms}$) emitted from the discharges. Detected lightning signals are in the Very High Frequency (VHF) to Very Low Frequency (VLF) range. Finally, lightning can be detected from space through its radiant energy.

Ground-based lightning systems are composed of multiple antennas, while mobile systems often deployed on aircraft use a direction and sense antenna. Various types of radio/magnetic direction-finding techniques can be applied (Cummins et al., 1998). To determine the distance of a discharge, fixed ground-based systems use

triangulation techniques, time-of-arrival (Shao & Krehbiel, 1996), or VHF techniques (Defer et al., 2001), while mobile systems use the signal frequency and its attenuation.

Since the mid 1980s, measurements from a variety of ground-based VLF/LF lightning detection networks (providing two-dimensional, 2-D, maps) with regional coverage are available such as the EUROpean Co-operation for LighTning Detection (EUCLID) in Europe (Schulz et al., 2016) and the National Lightning Detection Network (NLDN) in the contiguous United States (CONUS). These networks detect most CG flashes and a smaller percentage of IC flashes (Orville, 2008; Y. Zhu et al., 2020). The Earth Networks Total Lightning Detection Network (ENTLN) consists of wideband sensors that can separate CG and IC discharges (C. Liu et al., 2014; Rudlosky, 2015; Y. Zhu et al., 2017). ENTLN was initially focused on the U.S. but now has sensors in other regions of the world. The U.S. regional variability in LNO_x PE has been investigated through a comparison of ENTLN data with NLDN data (Lapierre et al., 2020) but also by using ENTLN data solely (Zhang et al., 2020).

A more sophisticated VLF/LF system is the Lightning Location Network (LINET), which is set up in smaller areas (~ 100 km radius). This system has the advantage of being able to detect low-current discharges (> 1 kA) and separate CG and IC flashes through refined antenna techniques and optimized waveform handling (Betz et al., 2004, 2009; Höller et al., 2009). Huntrieser et al. (2008, 2009, 2011) combined airborne in situ measurements of NO_x with LINET measurements to obtain the LNO_x PE for Brazilian, Australian, and West African thunderstorms.

VHF sensor networks such as the Lightning Mapping Array (LMA) system (Rison et al., 1999; Thomas et al., 2004) provide high temporally and spatially resolved measurements in three-dimensions (3-D) on a local to regional scale. Several LMA networks are deployed in regions of the U.S. During the Deep Convective Clouds and Chemistry (DC3) field experiment in the U.S., LMA data were used along with NO_x measurements to derive the LNO_x PE (Davis et al., 2019; Pollack et al., 2016). More recently, Pérez-Invernón, Gordillo-Vázquez, van der Velde, et al. (2023) and Pérez-Invernón, Gordillo-Vázquez, Huntrieser, and Jöckel (2023) used lightning data from the Ebro LMA network in Spain to estimate the regional LNO_x PE for this area.

The World Wide Lightning Location Network (WWLLN), which consists of VLF-sensors with a spacing of as much as 7,000 km, conducts global real-time measurements of flashes, mainly CG, with a low detection efficiency (ranging from $< 1\%$ to $\sim 20\%$). WWLLN has been in operation since 2003 (Hutchins et al., 2012; Jacobson et al., 2006; Lay et al., 2004; Rudlosky & Shea, 2013). Based on this data set, a global gridded data set, *WWLLN Global Lightning Climatology* (WGLC), has been established by Kaplan and Lau (2021). Lightning data from WWLLN have been used by Pickering et al. (2016), Bucsela et al. (2019) and Allen et al. (2019) to derive LNO_x PE for the Gulf of Mexico, the mid latitudes and the tropics, respectively.

A number of commercial VLF lightning networks also exist including the Global Position And Tracking Systems (GPATS) based in Australia that detects CG flashes (Kuleshov et al., 2011) and the Vaisala's Global Lightning Data set (GLD360) (Mallick et al., 2014). The latter network has been compared to the EUCLID system in Europe by Pohjola and Makela (2013). A further global lightning network is the Earth Networks Global Lightning Network (ENGLN) composed of both broadband sensors from ENTLN and VLF sensors from WWLLN. According to Marchand et al. (2019), ENGLN has a detection efficiency of 90% for CG strokes over the U.S. Data from both the ENGLN and EUCLID lightning networks has been used to derive the LNO_x PE over the Iberian Peninsula (Pérez-Invernón, Huntrieser, Erbertseder, et al., 2022; Pérez-Invernón, Huntrieser, Jöckel, & Gordillo-Vázquez, 2022).

The first satellite observations of lightning from space were conducted in the late 1970s. During the 1980s, space shuttle campaigns observed lightning using low-light television cameras. The Optical Transient Detector (OTD; 1995–2000; observing latitudes $\pm 75^\circ$) and the Lightning Imaging Sensor (LIS) on the TRMM satellite (1998–2015; observing latitudes $\pm 38^\circ$) detected CG and IC flashes from space from a Low Earth orbit (LEO) view on a daily basis. While a given location was only sampled briefly each day, a climatology for latitudes $\pm 55^\circ$ was obtained for OTD and $\pm 38^\circ$ for TRMM/LIS (Cecil et al., 2014; Christian et al., 2003; Mach et al., 2007). In addition, the International Space Station (ISS) was equipped with a LIS instrument from 2017 to 2023 (Blakeslee et al., 2020), which covered latitudes $\pm 55^\circ$. Miyazaki et al. (2014) and Marais et al. (2018) have used the combined OTD and TRMM/LIS data in combination with models to retrieve the LNO_x PE globally.

Based on combined OTD and TRMM/LIS satellite observations, a global-scale flash rate of 46 fl s^{-1} was derived by Cecil et al. (2014). This value has been updated to 59 fl s^{-1} by Cecil et al. (2026a, 2026b) after incorporation of

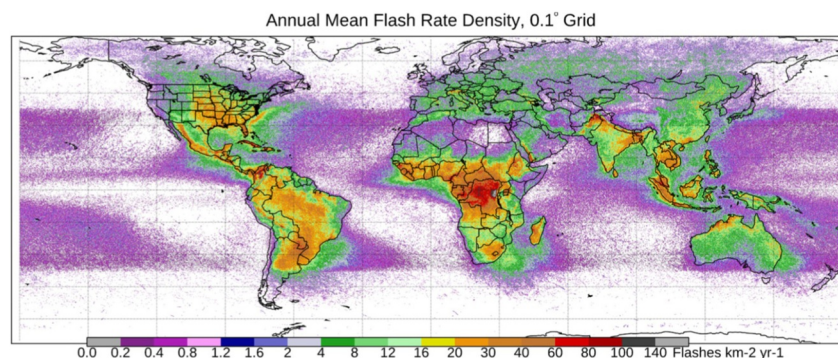


Figure 2. High resolution (0.1° grid) climatology of annual flash rate density from OTD (April 1995–March 2000), TRMM/LIS (January 1998–April 2015), and ISS/LIS (March 2017–November 2023) satellite instruments (from Cecil et al., 2026a). Copyright American Meteorological Society. Used with permission.

the ISS/LIS data in the global climatology and use of lower detection efficiency (DE) values than were previously assumed. The lower DE results from newer validation efforts (Virts et al., 2024) using more modern and higher quality ground-based detection systems that record many small and/or dim flashes not seen by the LIS instruments. Cecil et al. (2026a, 2026b) note that forthcoming use with LIS of the luminous event clustering algorithm (Mach, 2020) developed for use with the Geostationary Lightning Mapper (GLM) will likely reduce the 59 fl s^{-1} by 10%–15%. It is unknown whether the small and/or dim flashes produce significant NO_x , but because LNO_x production has been found to be dependent on flash size and flash energy (see Section 3), it is unlikely that they make a significant contribution. Therefore, the 46 fl s^{-1} flash rate from the Cecil et al. (2014) climatology is used throughout our manuscript to normalize the production rate of NO_x by lightning ($\text{LNO}_x \text{ PE}$) obtained by use of other instruments. In addition, the global and regional modeling studies from the literature that are presented in Section 3.5 use the older OTD/LIS climatology to scale parameterized flashes. Figure 2 shows the annual lightning climatology developed from the OTD, TRMM/LIS, and ISS/LIS observations (Cecil et al., 2026a). Prominent features include the land-sea contrast, the Amazon Basin-Congo contrast and the tropics-midlatitudes contrast. Blakeslee et al. (2020) obtained a global-scale flash rate based on 3-year of ISS/LIS data of $\sim 44 \text{ fl s}^{-1}$, which is the same rate obtained by Christian et al. (2003) using only OTD data. Both of these studies used the larger detection efficiencies available at those times. Studies on regional flash rate distributions have been conducted by Albrecht et al. (2016), Cecil et al. (2015), M. Peterson et al. (2021), and Cecil et al. (2026a) to define hot spot regions of high flash rates globally.

In November 2016 and March 2018, geostationary satellites were launched carrying GLM instruments designed for real-time mapping of total lightning. When positioned at $\sim 75^\circ\text{W}$ (GOES-East position) and $\sim 130^\circ\text{W}$ (GOES-W position), these satellites can sense flashes over most of the North and South American continents, the eastern Pacific and the western Atlantic Ocean (Goodman et al., 2013; Rudlosky et al., 2019). In December 2016, the Fengyun-4A Lightning Mapping Imager (LMI) (Hui et al., 2020; Ni et al., 2021; Yang et al., 2017) was launched. Its coverage range includes the Asian continent. Characteristics of the data set and information on how to request the data are available in Hui et al. (2023). In late 2022, the Meteosat Third Generation (MTG) Lightning Imager (LI), MTG-LI was launched. It measures lightning continuously over the European continent plus Africa and the eastern Atlantic Ocean (Erdmann et al., 2023; Holmlund et al., 2021). MTG-LI data were publicly released in July 2024 with the service declared operational on October 31 of that year. The advantage of such lightning observations from geostationary satellites is the continuous time sampling compared to low-orbiting satellites such as OTD and LIS (with passages only 1–2 times a day over a certain site). Allen, Pickering, Bucsela, et al. (2021) and Allen, Pickering, Lamsal, et al. (2021) used, for the first time, GLM data and separately ENTNLN data to determine the $\text{LNO}_x \text{ PE}$ over the U.S.

2.2. State of Knowledge as of 2007 Concerning LNO_x Production

Because of the great importance of LNO_x for atmospheric chemistry, several hundred scientific publications are available on this topic. To keep track of the wide knowledge on LNO_x , a first review was published in the 1990s by Lawrence et al. (1995). A decade later, a more extensive review on LNO_x and its quantification methods (as

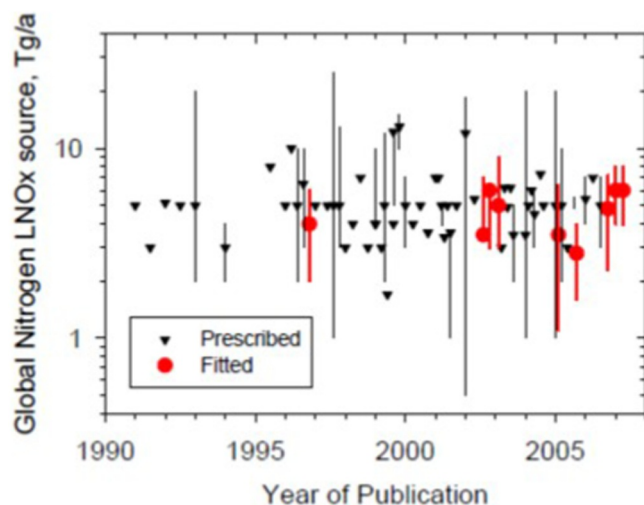


Figure 3. Time series showing global LNO_x source (Tg N yr^{-1}) as prescribed (triangles) or fitted (thick dots) in global atmospheric model studies as a function of year-of-publication. See Figure 28 of Schumann and Huntrieser (2007) for details.

given below) became available through Schumann and Huntrieser (2007). Now two decades later, our present review starts with a summary of those previous results on the LNO_x PE quantification.

In the Schumann and Huntrieser review, different methods for determining the LNO_x source were discussed. The methods can be divided into bottom-up and top-down approaches. The bottom-up approaches used theoretical calculations, laboratory and/or ground-based/airborne/satellite measurements of nitrogen oxides to derive the LNO_x production rate per flash (or storm). This number is then multiplied by the global number of flashes (storms) to obtain the global source. A range of $0.6\text{--}13 \text{ Tg N yr}^{-1}$, as given in Schumann and Huntrieser by considering many of these studies was obtained. This range is equivalent to:

- $2\text{--}40 \times 10^{25}$ NO molecules per flash (mean value of 15×10^{25} NO molecules per flash)
- $0.13\text{--}2.7 \text{ kg of N mass per flash}$ (mean value of $3.5 \text{ kg of N mass per flash}$)
- $30\text{--}660 \text{ mol NO}_x \text{ per flash}$ (mean values of $250 \text{ mol NO}_x \text{ per flash}$)

The large range is due to variations in flash type and flash location, which are still being investigated in detail. Findings by for example, Huntrieser et al. (2006) in Brazil, suggested that flashes in the midlatitudes produce more LNO_x per flash than flashes in tropical regions. Related to this also was the

question of whether CG flashes produce more LNO_x per flash than IC flashes, see for example, Barthe and Barth (2008).

To compare LNO_x PE values from different studies, it is essential to relate values to the same “standard” lightning detection system. Huntrieser et al. (2011) related all their LNO_x PE estimates based on regional lightning measurements from the ground-based Lightning Location Network (LINET) during different airborne field experiments (TROCCINOX in Brazil, SCOUT in Australia, and AMMA in Burkina Faso) to satellite-based global lightning measurements by the Lightning Imaging Sensor (LIS). Information on these lightning detection systems is given in Section 2.1. In this study, we will also relate values to LIS.

The storm-based extrapolation method requires an estimate of the number of storms occurring globally at any one moment ($\sim 1,800$ with a large uncertainty according to Ccopa et al., 2021). Since this estimate is more uncertain than estimates of the number of flashes occurring globally and the LNO_x PE derived after extrapolating from individual storms ($1\text{--}25 \text{ Tg N a}^{-1}$) is highly uncertain, the storm-based approach is rarely used today.

The top-down approaches adjust parameters in a global model in a way that best fits atmospheric and/or deposition measurements of various reactive nitrogen species. According to the review of a considerable number of such studies by Schumann and Huntrieser, accounting especially for observation-constrained model studies with different LNO_x parameterizations (see Figure 3), a best-fit estimate for the LNO_x source of $5 \pm 3 \text{ Tg N yr}^{-1}$ (or $2\text{--}8 \text{ Tg N yr}^{-1}$) was obtained. This range has become canonical and has been applied in many studies since 2007 (see Section 3). Still the uncertainty range is large and only slightly lower than the range estimated by Lawrence et al. (1995) in the first LNO_x review study ($1\text{--}8 \text{ Tg N yr}^{-1}$, with a best estimate of 5 Tg N yr^{-1}). The short review by Verma et al. (2021) also estimated the global LNO_x source to be $5 \pm 3 \text{ Tg N yr}^{-1}$. Though, to understand tropical tropospheric chemistry properly, Schumann and Huntrieser stated that an accuracy of $\sim 1 \text{ Tg N yr}^{-1}$ or 20% should be the goal. This criterion was obtained from a synopsis of detailed model results from different publications discussed in Section 2.10 and Table 15 in Schumann and Huntrieser (2007). While this goal has still not been reached, advances in understanding of physics and microphysics, technology, observing systems, and machine learning should make it achievable within the next 10–20 years. Perhaps it is time to start planning for a dedicated lightning- NO_x experiment that couples a field campaign with high-temporal resolution retrievals of NO_2 from geostationary sensors designed to handle high-reflectance scenes coupled with targeted observations of lightning and NO_x from miniature low-cost sensors with three-dimensional distributions of flash channels from satellite-borne lightning mapping arrays (Remington et al., 2024).

Schumann and Huntrieser (2007) also confirmed that extremely low and high values such as 0.2 Tg N yr^{-1} (Cook et al., 2000) and 220 Tg N yr^{-1} (Franzblau & Popp, 1989; Liaw et al., 1990), are unrealistic after accounting for measurements of atmospheric NO_x concentrations and nitrate deposition (Gallardo & Rodhe, 1997).

Operational lightning detection systems are incapable of detecting all discharge processes that produce lightning, which at the end may result in incorrect LNO_x PE values. For example, Rahman et al. (2007) found that undetectable (with common lightning detection schemes) slow discharges, such as quasi-steady continuing currents, may produce considerable amounts of LNO_x and suggested that they should be considered in global LNO_x estimates.

3. Production Efficiency (PE) of NO_x by Lightning

As more observations of lightning flashes and tropospheric NO_x have become available over the last two decades, a significant surge has occurred in the amount of research applying these data in analyses of LNO_x PE. There are five major techniques that have been employed to estimate PE:

- theoretical and laboratory work (Section 3.1),
- analysis of aircraft observations from storm anvils (Section 3.2),
- observation-constrained cloud-resolved modeling (Section 3.3),
- analysis of satellite observations (Section 3.4),
- global and regional modeling (Section 3.5).

The following subsections describe each technique and review the PE estimates derived from each method. The units of LNO_x PE vary between individual papers in literature. We have converted all units to mol fl^{-1} . Global model estimates of world-wide LNO_x production have been converted to mol fl^{-1} by dividing by the annualized equivalent of the OTD/LIS climatological satellite-derived estimate of $46 \text{ flashes s}^{-1}$ globally (Cecil et al., 2014). This climatology treats each flash as a point. It is important to note that data from a variety of lightning detection systems are used in making PE estimates. A flash defined by one system is not necessarily the same as one defined by another system, leading to uncertainties in PE estimates. Whenever possible, referencing the flash observations to the OTD/LIS climatology or to LIS observations during the period of interest helps to reduce uncertainty. The OTD/LIS data have also been climatologically analyzed in terms of flash extent density (M. Peterson et al., 2021), but the point-based climatology has been used in most PE estimates. LIS observations during overpasses of a particular region during field campaigns have also been used as a standard flash reference. Flashes observed by other systems (e.g., LMA, NLDN, ENTLN, GLM, etc.) have been adjusted using ratios of these flashes to LIS flashes, and the PE estimates have been adjusted to use the LIS-equivalent flashes.

3.1. Theoretical and Laboratory Considerations

In this category of LNO_x PE estimates, theoretical characteristics of flash components combined with laboratory experiment results are applied to observed flashes in a variety of ways. Table 1 summarizes the studies that have used this approach in estimating PE. Cooray et al. (2009) found that scaling up laboratory spark results using only return stroke flash energy is insufficient to estimate lightning flash PE. Many other processes such as leaders, continuing currents, and M components (short-lived pulses superimposed on the currents) also need to be considered. Cooray et al. (2009) considered these processes more important for LNO_x production than the return stroke. Considering all of these discharge processes, results showed similar LNO_x production by an IC and a CG flash for a similar channel length, leading to an estimate of total global production of 4 Tg N yr^{-1} , which translates to a mean PE of 196 mol fl^{-1} when 44 fl s^{-1} is assumed. Koshak, Peterson, et al. (2014) carried the use of these theoretical considerations, as well as laboratory results, farther in the development of the Lightning Nitrogen Oxides Model (LNOM). The LNOM was applied to the Northern Alabama Lightning Mapping Array (LMA) data for Augusts of 2005–2009. National Lightning Detection Network (NLDN) data were used to determine which LMA flashes were CG flashes. LNOM computes LNO_x for each 10-m flash segment using the Cooray equations and Y. Wang et al. (1998) laboratory results. In addition to return strokes, hot core stepped leaders, stepped leader corona sheaths, hot core dart leaders, K-changes, continuing currents, and M-components were considered to contribute to CG flashes. For IC flashes hot core flash leaders, flash leader corona sheaths, and K-changes were included. Vertical distributions of LNO_x production were obtained for both types of flashes.

Table 1
Summary of Theoretical/Laboratory/Flash Observation Studies to Determine LNO_x PE

Reference	Methods	LNO_x PE (mol fl ⁻¹)
Cooray et al. (2009)	Theoretical consideration of NO production by flash components	196; $PE_{IC} = PE_{CG}$ for similar channel length
Koshak, Peterson, et al. (2014)	Theoretical considerations for flash components combined with laboratory NO production results and Northern Alabama LMA flash characteristics in LNO _x model; analysis for August 2005–2009	$PE_{CG} = 484$ $PE_{IC} = 35$; assuming climatological mean IC/CG, PE = 147
Koshak (2014)	LNO _x model run with Northern Alabama LMA data for all of 2004–2012	$PE_{CG} = 604$ $PE_{IC} = 36$; assuming climatological mean IC/CG, PE = 178
Carey et al. (2016)	LNO _x model run with Northern Alabama LMA data for one storm from DC3 field program	$PE_{CG} = 919$ $PE_{IC} = 116$ Assuming climatological IC/CG, mean PE = 237
Davis et al. (2019)	Colorado and Northern Alabama LMA flash extent data used along with laboratory NO production rate	PE = 124 anomalously charged storms PE = 77 normally charged storms
Wu et al. (2023)	Optical energy from Geostationary Lightning Mappers over USA used along with laboratory NO production per unit energy and an assumed global mean production per flash of 250 mol fl ⁻¹	PE range: 468 winter 197 summer
Most likely value	From Koshak (2014)	178

Figure 4 illustrates the vertical distribution of LNO_x production for Northern Alabama LMA flashes during Augusts of 2005–2009. The mean LNO_x PE for CG flashes was found to be 484 mol fl⁻¹ while the mean PE for IC flashes was found to be 35 mol fl⁻¹. LNO_x does not include LNO_x production from continuing currents and associated M components for IC flashes. Therefore, the resulting IC PE is likely an underestimate. Koshak (2014) extended the LNO_x analysis to additional months and years (all months of years 2004–2012), finding mean PE for CG flashes to be 604 mol fl⁻¹ and mean PE for IC flashes to be 36 mol fl⁻¹. Carey et al. (2016) applied the LNO_x model to Northern Alabama LMA flash data for the 21 May 2012, multicell storm system observed during the DC3 field program. This analysis resulted in a mean PE for CG flashes of 919 and 116 mol fl⁻¹ for IC flashes. In comparison, using the climatological IC/CG flash ratio (Boccippio et al., 2001) for northern Alabama, a mean PE for all flashes in the region was found to be 237 mol fl⁻¹.

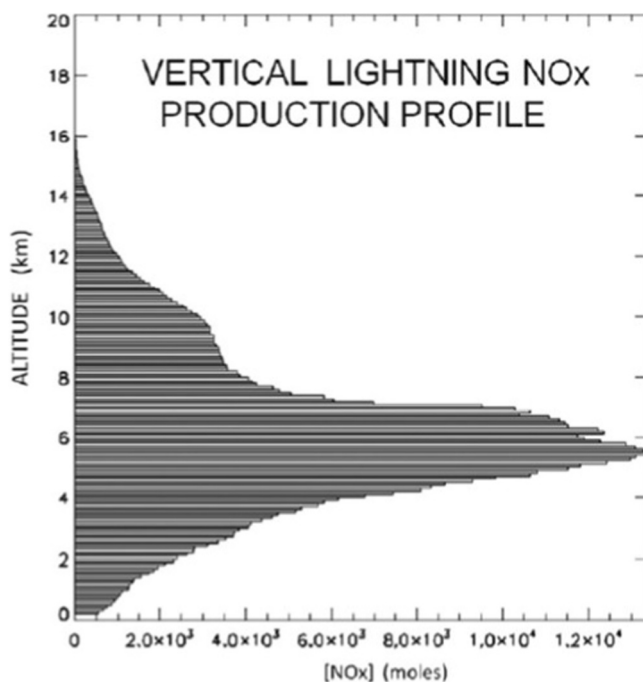


Figure 4. Vertical profile of LNO_x production computed from LNO_x model for the combination of CG and IC flashes observed by the Northern Alabama LMA (from Koshak, Peterson, et al., 2014).

Davis et al. (2019) used the laboratory results of Y. Wang et al. (1998) for LNO_x production in estimating PE for flashes in five storms observed by LMAs in Colorado and Alabama during DC3. Both anomalously charged (positive charge in mid-troposphere) and normally charged (negative charge in mid-troposphere) storms were included. Radar-derived winds were used to advect air parcels from 1-km thick parcels specified by LMA flash extent data. Storm mean PE ranged from 72 to 158 mol fl⁻¹. Anomalous storms had a mean of 124 mol fl⁻¹, while normal storms had a mean of 77 mol fl⁻¹. Anomalously charged storms tend to have greater flash rate and stronger updrafts, and therefore greater LNO_x in the upper troposphere. Typically, larger flash rates result in smaller PE values (e.g., Allen et al., 2019; Bucsela et al., 2019; Koike et al., 2007), but in the anomalously charged cases studied by Davis et al. (2019) this result was not found.

Wu et al. (2023) used GLM observations and the so-called Beta (β) approach (Koshak, Vant-Hull, et al., 2014) to model flash-to-flash variations in PE (Equation 1). Y , the NO_x thermochemical yield was assumed to equal 10^{17} molecules NO J⁻¹ based on analysis conducted by Borucki and Chameides (1984) of laboratory results. Q is the flash optical energy observed by GLM. β , the fraction of the total flash energy seen by a satellite, was derived for GLM by dividing the optical energies observed over an extended period of time (e.g., 1 year) by an assumed total energy for a flash (~ 1 GJ) obtained

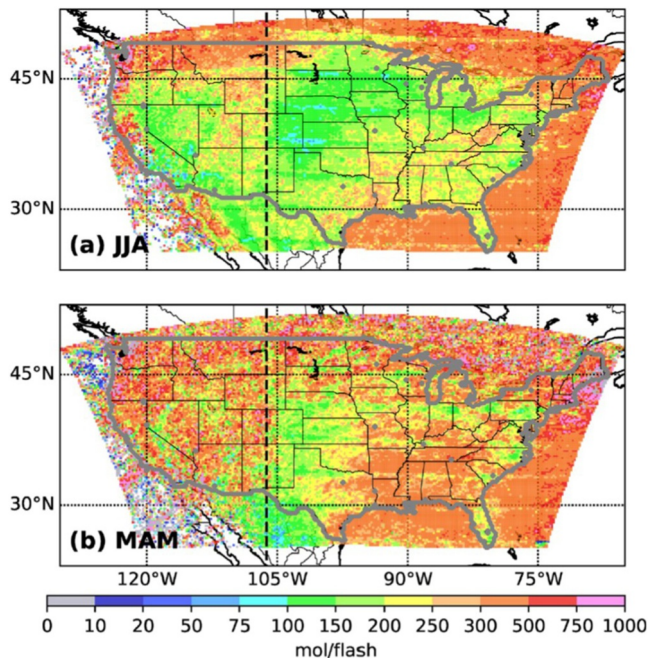


Figure 5. Average LNO_x per flash estimated from 2019 to 2021 GLM observations of flash optical energy over CONUS for (a) June, July, August; (b) March, April, May. From Wu et al. (2023).

from previous laboratory work (Borucki and Chameides, 1984), while N_a is Avogadro's number.

$$\text{PE} = (Y Q)/(\beta N_a) \quad (1)$$

Thus, the PE for an individual flash is proportional to Q . After choosing β so that the mean LNO_x PE over the CONUS for the entire year equals 250 mol per GLM flash, the mean PEs for DJF, MAM, JJA, and SON were found to be 468, 265, 197, and 267 mol fl^{-1} , respectively. The smaller PE in JJA is associated with smaller optical energy per flash likely due to greater flash rates. The geographic distribution of mean PE over CONUS for JJA and MAM is shown in Figure 5.

H. S. Peterson and Beasley (2011) postulated a mechanism by which ice crystals within the lightning discharge channel adsorb N atoms generated by lightning flashes. The adsorption extends the lifetime of N atoms making them more likely to be available for reactions. They conducted chemical calculations at a range of temperatures and found that reactions catalyzed by ice crystals could produce 2.7 times more NO than reactions without the ice crystal catalysis.

Koshak (2014) provide the most comprehensive analysis using a combination of theoretical and laboratory results applied to lightning observations for a full 9 years, and therefore this work is likely the most useful in defining PE by this method. Assuming a climatological mean IC/CG flash ratio of 3, the overall LNO_x PE for the analysis period is 178 mol fl^{-1} . The other values in Table 1

have not been used in defining a best estimate for this category for a variety of reasons. The Cooray et al. (2009) value shown in Table 1 has been updated to assume a 44 flash per second flash rate, but it is still not used because it assumes an uncertain 45 km channel length for both IC and CG flashes. The period of record considered by Koshak, Peterson, et al. (2014) is a subset of that in Koshak (2014). The Carey et al. (2016) analysis is for only one storm. We do not know the fraction of Colorado and Alabama storms that are anomalously charged, preventing us from obtaining a valid overall estimate to use from the Davis et al. (2019) work. Wu et al. (2023) assume 250 mol fl^{-1} for the mean global PE as the starting point in their analysis, thereby yielding results that are relative to that value.

3.2. Analysis of Aircraft Data Observed in Storm Anvils

There have been five airborne field experiments in the last 18 years with an objective to study the impact of lightning on NO_x production. These experiments typically involve transects of thunderstorm anvils or flight segments downwind of storms. PE is determined using the in situ observations of anvil NO_x to compute the LNO_x flux or together with the storm volume to estimate the anvil LNO_x content. This information is used along with observed flash rates to estimate PE. Table 2 summarizes the use of aircraft data for estimating PE.

Koike et al. (2007) reported LNO_x production in storms over/near the Gulf of Carpentaria in northern Australia during the Biomass Burning and Lightning Experiment—Phase C (BIBLE-C). Gulfstream II aircraft observations were conducted at altitudes of 11.5–14 km, 10–14 hr transport time downwind of these storms. NO was observed using a chemiluminescent detector, and NO_2 was estimated from the photostationary state. Lightning flashes from the Global Positioning and Tracking System (GPATS) over Australia were available to allow estimates of flash rate and of PE. The more intense storms (those with larger flash rates) observed on 8–9 December 2000, yielded 32–73 mol fl^{-1} , whereas the weaker storms of 11–12 December 2000, had much greater PE of 349–818 mol fl^{-1} . A range of LNO_x PE values is shown for each storm to account for uncertainties in the GPATS CG flash detection. Total flashes were estimated from observations of CG flashes using the cold-cloud depth parameterization of Price and Rind (1993).

The German DLR Falcon aircraft was deployed in several field missions to sample deep convective anvils in tropical and subtropical environments. Huntrieser et al. (2007, 2008, 2009, 2011) report on in situ trace gas observations from these missions and made estimates of LNO_x production. The Tropical Convection, Cirrus, and

Table 2

Summary of Analyses of Aircraft Data for Estimating LNO_x PE

Reference	Experiment	Location	Flash obs. system	LNO_x PE (mol fl ⁻¹)	LNO_x PE (mol fl ⁻¹) adjusted to LIS equivalent flashes
Koike et al. (2007)	BIBLE-C	Australia	GPATS	32–73 intense storms; 349–818 weaker storms	N/A
Huntrieser et al. (2008)	TROCCINOX	Brazil	LINET	–	49 tropical 98–148 subtropical
Huntrieser et al. (2009)	SCOUT-O3	Australia	LINET	–	292–343 Hector 386 subtropical 121 tropical
Huntrieser et al. (2011)	AMMA	West Africa	LINET	–	71 south of ITCZ 176 north of ITCZ
Pollack et al. (2016)	DC3	Oklahoma	LMA	For most comprehensively sampled cases: 141–205 anvil flux; 107–292 anvil volume	154–224 anvil flux; 117–320 anvil volume
Pollack et al. (2016)	DC3	Colorado	LMA	For most comprehensively sampled case: 131 anvil flux 143 anvil volume	N/A
Mean ± sigma (most likely range)					192 ± 106

Nitrogen Oxides (TROCCINOX) experiment was conducted over southern Brazil in February and March 2004 and February 2005. Both tropical and subtropical convective systems (anvil region up to 12.5 km) were sampled by the aircraft. Lightning strokes were recorded by a Lightning Location Network (LINET) system in the 2005 experiment. The LNO_x flux out of the anvil was estimated from the mean observed NO_x mixing ratio in the outflow, the outflow wind velocity, and the anvil cross sectional area (Huntrieser et al., 1998). The flux was then related to the LINET strokes (Huntrieser et al., 2008) and the LINET strokes were scaled to Lightning Imaging Sensor (LIS) flashes. Huntrieser et al. (2008) found a PE of 49 mol fl⁻¹ in tropical convection and 98–148 mol fl⁻¹ in subtropical systems. It was hypothesized that the greater PE in the subtropical systems was a consequence of greater flash lengths associated with their larger vertical wind shears. See following sections.

Huntrieser et al. (2009) reported on the SCOUT-O3/ACTIVE mission, based in Darwin, Australia in November–December 2005. A LINET system was installed to record strokes in the vicinity of Darwin and the offshore Tiwi Islands. The strokes were again scaled to LIS flashes to allow comparison of LNO_x PE to other studies. Nearly daily mesoscale convective systems, that develop in response to sea breeze convergence over the Tiwi Islands, are named Hector storms. Their regular occurrence simplified the airborne sampling. NO , O_3 , and j (NO_2) were measured on the Falcon, and NO_2 was estimated from the photostationary state assumption. Based on the LNO_x flux in the anvil up to 12.5 km and scaled flashes, the Hector storm sampled on 19 November 2005, yielded a PE of 292–343 mol fl⁻¹. Among the observed storms in Australia, the Hector system contained the greatest stroke rate. Continental subtropical systems south of Darwin had a PE of 386 mol fl⁻¹, while continental tropical systems near Darwin had a smaller PE of 121 mol fl⁻¹. Wind shear may play a role in these differences too (see Figure 6). One component of the African Monsoon Multidisciplinary Analysis (AMMA) field phase in August 2006 was airborne atmospheric chemistry sampling of the deep convection over West Africa. The Falcon was based in Ouagadougou, Burkina Faso. LINET strokes were recorded from stations in Burkina Faso and Benin, and the recorded strokes were scaled to LIS flashes. Huntrieser et al. (2011) presented results for a mesoscale convective system (MCS) south of the Inter-Tropical Convergence Zone (ITCZ) with PE of 71 mol fl⁻¹ and for a MCS north of the ITCZ with PE of 176 mol fl⁻¹, based on the anvil LNO_x fluxes (up to 12 km) and LIS-equivalent flashes.

Pollack et al. (2016) analyzed in situ NO_x data from the NASA DC-8, NSF/NCAR GV, and DLR Falcon aircraft flights through convective anvils at 10.1–12.3 km over Oklahoma (5 storms) and Colorado (3 storms) during the DC3 experiment in May/June 2012. NO measurements were made via chemiluminescence on all three aircraft, and NO_2 was mainly measured using UV photolysis followed by chemiluminescence detection. Two methods of analysis to arrive at LNO_x PE estimates were employed: anvil volume and anvil flux methods. Radar and satellite images were used to estimate the volume of the anvil into which LNO_x had been injected, and the observed NO_x enhancement was assumed to be equally distributed within this volume. The anvil flux method followed the

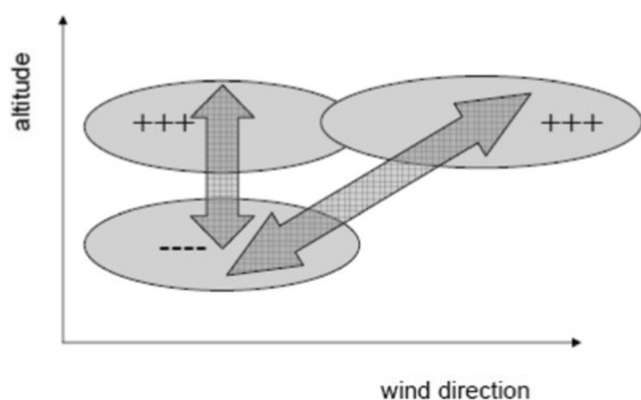


Figure 6. The two-way arrows show the flash channel length for an idealized storm with no vertical wind shear (vertical arrow) and a storm with substantial wind shear (diagonal arrow). The arrows illustrate the hypothesis that enhanced vertical wind shear can lead to greater horizontal separation of primary charged regions, leading to greater flash lengths and greater LNO_x production per flash. From Huntrieser et al. (2008).

approach of Huntrieser et al. (1998). Lightning Mapping Arrays (LMAs) were deployed in both regions, and flashes were constructed from the LMA VHF sources. For the three storms with the most comprehensive sets of measurements the PE results from the anvil volume and anvil flux methods, respectively, are: 29 May 2012 Oklahoma severe MCS, 141 and 107 mol fl^{-1} ; 16 June 2012 Oklahoma multicell, 205 and 292 mol fl^{-1} ; and 22 June 2012 Colorado supercells, 131 and 143 mol fl^{-1} . After analysis, the volume method was judged to be more robust in this analysis. These PE values are similar to or greater than those found by Davis et al. (2019) using the combination of laboratory LNO_x production results and LMA data. Mean storm LNO_x PE for the Oklahoma and Colorado storms was found to be positively correlated with mean storm flash area observed by the LMAs (see Figure 7).

Labrador et al. (2009) analyzed upper tropospheric aircraft data from the portion of the ACTIVE campaign that occurred during the Northern Australia monsoon season. The NO_x background upwind of a storm southwest of Darwin with only a few flashes was relatively large (723 pptv), which was only ~ 250 pptv less than in the storm. Five-day back trajectories indicated that the air had a source over the convectively active northern Australian continent. These factors suggested that there is an accumulation of LNO_x over this region during the monsoon season.

3.3. Cloud-Resolved Modeling

The methods for determining LNO_x PE described in Section 3.2 directly used aircraft observations from storm anvils. However, these data represent only a small fraction of the volume of a thunderstorm cloud. Much volume of the anvil, as well as the convective core region, of each of the storms were not sampled, thereby leading to significant uncertainties in the resulting PE values. Cloud-resolving storm and chemistry models provide a fuller depiction of the 3-D LNO_x field within a storm and are most useful when constrained by anvil NO_x observations and observed flashes. DeCaria et al. (2000, 2005) and Fehr et al. (2004) pioneered this approach. Since then, several different cloud-resolving models have been employed in simulations of observed storms from a variety of

field projects. The models include a specification of flash number (both IC and CG), flash PE, and LNO_x emission distribution both in the horizontal and vertical. Typically, different PE values are tested until the best match of resulting model anvil NO_x mixing ratios and observed NO_x is obtained. Table 3 summarizes cloud-resolved modeling studies employed in estimating PE.

Barthe and Barth (2008) simulated the STERAO-A 10 July 1996, Colorado severe storm (Dye et al., 2000) with the WRF-AqChem model (Barth et al., 2007) using a new flash rate parameterization based on the multiplicative product of non-precipitation ice and precipitation ice and a filamentary LNO_x source placement scheme. This storm had been sampled by the University of North Dakota Citation aircraft and lightning flash data were available from a VHF interferometer. A PE value of $121 \pm 41 \text{ mol fl}^{-1}$ was obtained.

Ott et al. (2010) also simulated severe storms from STERAO-A (10 July and 12 July 1996, events). In addition, she simulated a supercell storm from the EULINOX project (Huntrieser et al., 2002) near Munich, Germany that had been sampled by the DLR Falcon, two storms from the CRYSTAL-FACE mission (Ridley et al., 2004) in south Florida, sampled by the NASA WB-57, and the June 11 squall line from the PRESTORM project over Oklahoma in 1985, for which no chemical sampling had been performed. The Goddard Cumulus Ensemble Model (GCE, 3-D) was used to simulate the July STERAO-A, EULINOX, and PRESTORM systems, and

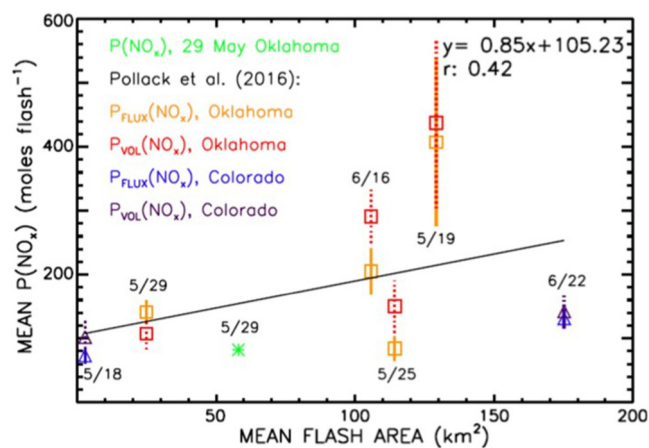


Figure 7. Scatterplot showing the correlation between the mean flash area and mean LNO_x PE (from Pollack et al., 2016) for six convective events observed during DC3. The squares and triangles represent Oklahoma and Colorado convection, respectively. LNO_x production using the flux-based approach is represented by orange and blue, while production by the volume-based approach is represented by red and purple. The green asterisk represents the 29–30 May event based on data within the 10-min moving spatial masks and 3D coverage of the central OK LMA network (23:40–03:00 UTC). Fitted line and correlation are for all points plotted. Plot from Pickering et al. (2024).

Table 3
Summary of Cloud-Resolving Model Studies Applied for Estimating LNO_x PE

Reference	Experiment/Storm date	Location	Model	Flash Obs. System	LNO_x PE (mol fl^{-1})	LNO_x PE (mol fl^{-1}) adjusted to LIS equivalent flashes
Barthe and Barth (2008)	STERAO-A—10 September 1996	Colorado	WRF-AqChem with filamentary lightning	ONERA interferometer	121 ± 41	N/A
Ott et al. (2010)	STERAO-A—10 September 1996	Colorado	GCE/CSCTM	ONERA interferometer	$\text{PE}_{\text{CG}} = 390$ $\text{PE}_{\text{IC}} = 234$	N/A
Ott et al. (2010)	STERAO-A—12 September 1996	Colorado	GCE/CSCTM	ONERA interferometer	$\text{PE}_{\text{CG}} = 460$ $\text{PE}_{\text{IC}} = 460$	N/A
Ott et al. (2010)	EULINOX—21 July 1998	Germany	GCE/CSCTM	BLIDS and ONERA interferometer	$\text{PE}_{\text{CG}} = 360$ $\text{PE}_{\text{IC}} = 414$	N/A
Ott et al. (2010)	CRYSTAL-FACE—29 July 2002	Florida	MM5/CSCTM	NLDN	$\text{PE}_{\text{CG}} = 590$ $\text{PE}_{\text{IC}} = 354$	$\text{PE}_{\text{CG}} = 525$ $\text{PE}_{\text{IC}} = 315$
Ott et al. (2010)	CRYSTAL-FACE—16 July 2002	Florida	ARPS/CSCTM	NLDN	$\text{PE}_{\text{CG}} = 700$ $\text{PE}_{\text{IC}} = 525$	$\text{PE}_{\text{CG}} = 623$ $\text{PE}_{\text{IC}} = 467$
Cummings et al. (2013)	SCOUT-O ₃ /ACTIVE	Australia	WRF-AqChem	LINET	500–600	320–384
Pickering et al. (2024)	DC3—29–30 May 2012	Oklahoma	WRFChem	Oklahoma LMA	80–110	88–120
Mean ± 1 sigma (most likely range)						355 ± 174

chemistry was run offline using the University of Maryland Cloud Scale Chemical Transport Model (CSCTM; DeCaria et al., 2005). The subtropical and tropical convective systems sampled in CRYSTAL-FACE were simulated using the Mesoscale Model version 5 (MM5) and the Advanced Regional Prediction System (ARPS) meteorological models, respectively. CSCTM simulations with offline chemistry and transport were also conducted for these events. Total lightning flashes were observed by a VHF interferometer during the STERAO-A and EULINOX storms while CG flashes were observed during CRYSTAL-FACE and then scaled to obtain total flashes using a modified climatological IC/CG ratio. The best match of model and aircraft-observed NO_x was determined for each storm after testing an array of possible PE values. The resulting PE values for each storm are displayed in Figure 8. Averaged over the simulations of five storms from the three field experiments the mean CG PE equaled 500 mol fl^{-1} with a PE_{IC} to PE_{CG} ratio of 0.93. Ott et al. (2010) also developed vertical profiles of LNO_x production after post-convection mixing that have been used in mesoscale and coarser resolution models including the NASA Global Modeling Initiative (Allen et al., 2010).

Cummings et al. (2013) performed simulations of the Hector storm of 16 November 2005, observed during the SCOUT-O₃/ACTIVE campaign based in Darwin, Australia. A LINET system recorded lightning strokes. The 3-D WRF-AqChem model containing online chemistry and an LNO_x placement scheme was used along with the observed flashes. Simulated upper tropospheric NO_2 partial columns in this storm were considerably greater than most satellite observations of columns in tropical marine convective events (e.g., Bucsela et al., 2010), as tropical island convection such as Hector is more vigorous, reaches higher altitudes (20 km), and likely has longer channel lengths leading to more production of LNO_x . After constraining the model with Egrett aircraft NO_x observations, PEs of 500–600 mol of NO_x per LINET flash (320–384 mol per LIS flash) were found. Huntrieser et al. (2009) found that Hector storms had larger PE (292–343 mol per LIS flash) than other tropical Australian storm systems. The Cummings et al. model-based PE range overlaps this aircraft-based estimate.

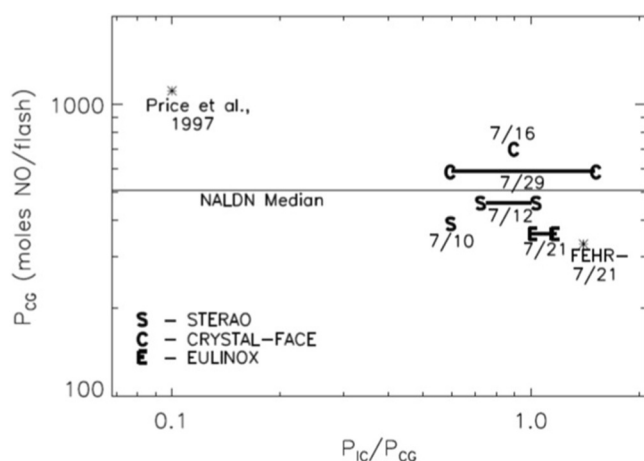


Figure 8. LNO_x production scenarios that best fit observations for the 16 July CRYSTAL-FACE (C, 7/16), 29 July CRYSTAL-FACE (C, 7/29), 10 July STERAO (S, 7/10), 12 July STERAO (S, 7/12), and 21 July EULINOX (E, 7/21) storms. The bars in the 7/29 CRYSTAL-FACE, 7/12 STERAO, and 7/21 EULINOX storms indicate the estimate uncertainty. The Price et al. (1997) and Fehr et al. (2004) production scenarios are indicated by asterisks, while the estimated value of P_{CG} calculated assuming the North American Lightning Detection Network median peak current from Orville et al. (2002) is indicated by the horizontal line. From Ott et al. (2010).

Pickering et al. (2024) obtained LNO_x PE values of 80–110 mol per LMA flash (88–120 mol per LIS flash) for a DC3 storm observed over northwestern Oklahoma on 29–30 May 2012. The simulation was run online in WRF-Chem (Cummings et al., 2024) using a vertical distribution of LNO_x consistent with observed flash channel segments for CG and IC flashes, and model flash rates obtained with the upward ice flux parameterization (Finney et al., 2014). The model NO_x was constrained by observations from the NASA DC-8 and NCAR GV during transects of the storm anvil. The results were in good agreement with the Pollack et al. anvil volume-based estimate of 117 mol fl^{−1}. The relatively small PE was associated with a high flash rate and relatively small mean flash extent, a finding that agrees with the satellite-based studies discussed in Section 3.4, and LMA flash analyses (e.g., Bruning & Thomas, 2015). In contrast, an Oklahoma line of severe convection observed on 19–20 May 2012, covered roughly a similar area to the May 29–30 storm, but had a much smaller flash rate. Pollack et al. (2016) found a much larger PE (436 mol fl^{−1} using anvil volume) for this storm, again suggesting agreement with LMA-based findings with regard to flash rate and size relationships.

Sato et al. (2023) used the Scalable Computing for Advanced Library and Environment (SCALE) model with an explicit bulk lightning model with offline chemistry to simulate LNO_x for a case in which enhanced NO_y was measured at the top of Mt. Fuji, Japan. The model computes lightning frequency from the electric field determined from charge density but does not

distinguish between IC and CG flashes. Assuming an LNO_x PE of 250 mol fl^{−1}, the best agreement between modeled and observed NO_y at Mt. Fuji was obtained using a two-mode vertical distribution of LNO_x, that is, an emission profile with a small peak near the surface and a larger peak at 11–12 km.

3.4. Satellite Data Analysis

The portion of the satellite-retrieved NO₂ column attributable to lightning is derivable from UV-VIS retrievals over deep convective clouds. Satellite observations allow quantification of the LNO_x content of a multitude of storms over large areas of the earth. Initial efforts to retrieve LNO_x columns using coarse-resolution retrievals from the GOME instrument were included in the Schumann and Huntrieser (2007) LNO_x review. Successive instruments (SCIAMACHY, OMI, TROPOMI) have progressively finer spatial resolution, which is useful given the spatial scale of thunderstorms. However, these instruments only provide one observation per day at a given location, and the overpasses are in the late morning or early afternoon when lightning activity is not at a maximum. The general approach for obtaining LNO_x estimates for deep convective pixels begins with subtraction of the stratospheric slant column of NO₂ from the total slant column to obtain a tropospheric slant column, which is then divided by an air mass factor (AMF) appropriate for deep convection to obtain a vertical column abundance of NO₂ affected by LNO_x production (e.g., Pickering et al., 2016). The AMF must be computed using a profile of NO₂ typical of a deep convective storm containing LNO_x production, which has a peak in NO₂ in the middle to upper troposphere. Profiles typically come from a model simulation. The models employed for obtaining appropriate profiles are mentioned in the following paragraphs. The resulting NO₂ column amount is multiplied by model-derived NO_x/NO₂ ratios appropriate for time of day and latitude and longitude of the pixel to obtain the NO_x column amount. A background column amount must then be subtracted to remove NO_x resulting from boundary layer pollution and/or NO_x from upwind lightning, thus obtaining the LNO_x for the pixel. A variety of methods have been used to estimate the background. The specific method for each satellite application is given in the paragraphs below. Table 4 summarizes the use of remote sensing data for estimating LNO_x PE.

Beirle et al. (2009) simulated convective systems over the Pacific Ocean with a cloud-resolving chemistry model and used the resulting hydrometeor mixing ratios in a radiative transfer model to compute box AMFs. These AMFs were used along with model NO₂ profiles to compute slant column NO₂ amounts. They found a mean ratio

Table 4
Summary of Use of Remote Sensing Data in Estimating LNO_x PE

Reference	Instrument	Region	Flash Obs. System	LNO _x PE (mol fl ⁻¹)	LNO _x PE (mol fl ⁻¹) adjusted for LIS equivalent flashes
Beirle et al. (2010)	SCIAMACHY	Global	WWLLN scaled to OTD/LIS	—	Majority <17 Some 42–63
Bucsela et al. (2010)	OMI	Central America; Tropical Pacific	WWLLN and Costa Rica Networks scaled to LIS	—	100–250
Pickering et al. (2016)	OMI	Gulf of Mexico	WWLLN scaled to OTD/LIS	—	80 ± 45
Bucsela et al. (2019)	OMI	Northern mid-latitude continents	WWLLN scaled to OTD/LIS	—	180 ± 100 Smaller PE found with larger flash rate
Allen et al. (2019)	OMI	Tropics	WWLLN scaled to OTD/LIS	—	170 ± 100 150 continental 342 marine
Seiler et al. (2025)	OMI	Central US; Central America; West Africa	WWLLN scaled to LIS	52–80 Central US—WWLLN only	439–691 marine Central America; 205–431 West Africa
Zhang et al. (2020)	BEHR OMI	CONUS	ENTLN	—	90 ± 50
Lapierre et al. (2020)	BEHR OMI	CONUS regions	ENTLN and NLDN	—	24
Allen, Pickering, Bucsela, et al. (2021)	TROPOMI	CONUS and nearby ocean	GLM and ENTLN	GLM: 175 ± 100 ENTLN: 120 ± 65	GLM: 112 ± 64 ENTLN: 154 ± 83
Pérez-Invernón et al. (2022a)	TROPOMI—KNMI and DLR	Pyrenees and Ebro Valley, Spain	ENGLN and EUCLID	—	47–69
Pérez-Invernón et al. (2023a)	TROPOMI	Ebro Valley, Spain	LMA	—	54–108
Zhang et al. (2023)	TROPOMI	Arctic	GLD360	Estimated moles LNO ₂ per stroke; ocean PE 6X larger than land PE	—
Allen, Pickering, Lamsal, et al. (2021)	GCAS	Alabama, Colorado, Gulf of Mexico, Atlantic	GLM and ENTLN	GLM: 360 ± 180 ENTLN: 230 ± 115	GLM: 194 ± 97 ENTLN: 294 ± 147
Mean ± 1 sigma (most likely range)					187 ± 161

of slant column NO₂ to vertical column NO_x of 0.46, indicating that NO₂ from lightning should have good visibility from space. Then, Beirle et al. (2010) estimated LNO_x PE using detection-efficiency adjusted flash counts from the World Wide Lightning Location Network (WWLLN) and 30 × 60 km late-morning NO₂ columns from SCIAMACHY. Focusing on pixels over storms in low-pollution areas with flash rates exceeding 1 km⁻² in the hour prior to satellite overpass, they obtained PE values of less than 17 mol per flash for most storms with slightly higher values (42–63 mol per flash) for a few events over/near the coast of North America.

Bucsela et al. (2010) were the first to use the finer resolution OMI instrument (13 × 24 km pixels at nadir) on the NASA Aura satellite for estimating LNO_x PE. This work used WWLLN flashes with the detection efficiency determined using the Costa Rica ground-based network and the TRMM Lightning Imaging Sensor (LIS). Four storms were analyzed during the NASA Tropical Composition, Cloud, and Climate Coupling (TC⁴) field program, based in San Jose, Costa Rica, using pixels over storms and in-storm outflow. A priori NO₂ profiles were derived from NASA DC-8 aircraft observations in these regions. Background was obtained from OMI columns on

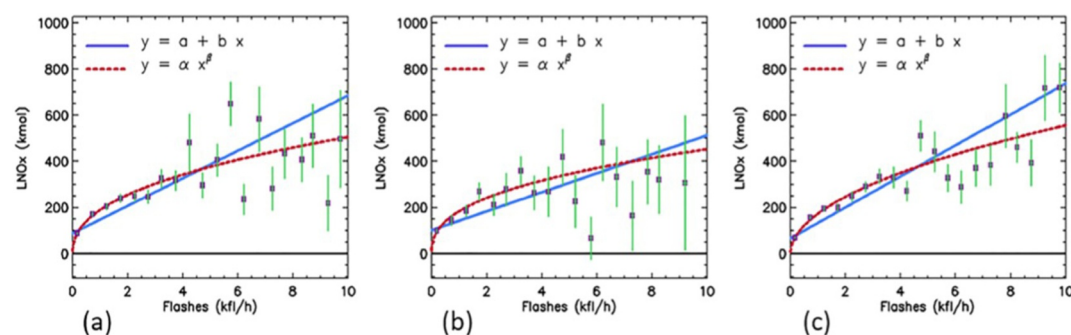


Figure 9. LNO_x (10^3 mol) versus 1-hr flashes (10^3) using data binned by flash count for (a) North America, (b) Europe, and (c) East Asia. Blue and red lines are linear and power-function fits to the data, respectively. Vertical bars are standard errors. From Bucsel et al. (2019).

5 days during July–August 2007 with minimal convective activity in the TC^4 region. The resulting PE over the four storms ranged from 100 to 250 mol fl^{-1} .

Pickering et al. (2016) used detection-efficiency adjusted WLLN flashes and OMI retrievals over deep convective pixels (i.e., pixels with cloud radiance fraction > 0.9 and optical centroid pressure $< 650 \text{ hPa}$) to estimate LNO_x PE for a multitude of storms over the Gulf of Mexico during June–August of 2007–2011. By subtracting a NO_2 background based on aircraft flight data from the Gulf of Mexico from the OMI retrieval, lightning- NO_2 columns were obtained.

Profiles of LNO_2 and LNO_x derived from “no-lightning” and “lightning” simulations with the NASA Global Modeling Initiative (GMI) model were used to convert the lightning- NO_2 columns derived from OMI to NO_x columns. PE was calculated using summation and regression methods. In the summation approach, the PE was obtained by dividing the total moles of LNO_x observed over the region during the month/season by the total detection-efficiency adjusted flashes over the same area and time. In the regression approach, the PE was assumed to be given by the slope of a linear fit to a scatterplot of daily moles of LNO_x and daily flashes over the region. The resulting mean PE and uncertainty using the two approaches was $80 \pm 45 \text{ mol fl}^{-1}$. To ensure that the LNO_x PE estimate was based on a significant LNO_x signal from a given pixel, a flash threshold of 3,000 flashes per $1^\circ \times 1^\circ$ grid box over the 3 hr prior to OMI overpass was employed.

Subsequent OMI analyses were expanded to encompass much larger regions. Bucsel et al. (2019) analyzed OMI LNO_x and WLLN lightning data (scaled using OTD/LIS) over the Northern Hemisphere mid-latitude continents. A priori profiles from the GMI model (as in Pickering et al. (2016)) were used in the specialized AMFs needed for these calculations. In this analysis no flash threshold was used, and background was estimated from OMI deep convective pixels that contained no flashes. Mean PE and uncertainty over these continental regions and over five boreal summers was found to be $180 \pm 100 \text{ mol fl}^{-1}$. Bucsel et al. (2019) found that a power function is a better fit to the relationship between flash rate and LNO_x mass than a linear fit indicating an inverse dependence of PE on flash rate (Figure 9), consistent with the inverse relation between flash rate and flash size from LMA data (e.g., Bruning & Thomas, 2015).

Allen et al. (2019) conducted an analysis of OMI LNO_x over the tropics, also using $1^\circ \times 1^\circ$ GMI a priori profiles and WLLN lightning data (scaled to OTD/LIS). This analysis was conducted over both continental and marine regions of the tropics and over both the austral and boreal summers of 2007–2011 using linear regression between flashes and OMI LNO_x . The y-intercept of the regression was considered to represent the background. The mean PE was found to be $170 \pm 100 \text{ mol fl}^{-1}$, very similar to that found by Bucsel et al. (2019) for mid-latitude continents. In the tropics, the marine regions where flash rates were usually low had a PE a factor of two larger than in the tropical continental regions where flash rates were usually high. The high uncertainty range in the OMI-based estimates ($> 50\%$) is a concern but is typical for these types of estimates. Uncertainty in the LNO_x retrieval itself was estimated at $\sim 36\%$. Seiler et al. (2025) compared OMI LNO_x columns within three storms with LNO_x columns estimated from in-cloud aircraft data and found comparable uncertainties. Seiler et al. also

estimated PE for the storms using OMI data. The June 11 DC3 mesoscale convective system had a very high flash rate and a low PE ($52\text{--}80\text{ mol fl}^{-1}$). The 5 August 2007, TC⁴ tropical marine convection had a relatively low flash rate and a large PE ($433\text{--}691\text{ mol fl}^{-1}$). The 6 August 2006, AMMA continental tropical storm had an estimated PE of $205\text{--}431\text{ mol fl}^{-1}$, roughly a factor of two smaller than the TC4 storm. This result is consistent with the Allen et al. (2019) finding of PE in tropical continental regions having a mean PE approximately a factor of two smaller than in tropical marine regions.

Two studies have used the Berkeley High Resolution (BEHR) OMI product for LNO_x PE estimation. The BEHR retrievals start with the same slant columns but use higher-resolution (12-km) a priori profiles from WRF-Chem as the basis for the shape functions needed to convert the slant columns to vertical columns. Zhang et al. (2020) performed an analysis for summer 2014 over the contiguous United States (CONUS), using Earth Networks Total Lightning Network (ENTLN) flashes. Background was estimated from WRF-Chem grids over clean and polluted regions. A mean PE and uncertainty of $90 \pm 50\text{ mol fl}^{-1}$ was obtained. The Lapierre et al. (2020) analysis used the BEHR OMI product with similar a priori profiles as Zhang et al. (2020) and a combination of ENTLN and National Lightning Detection Network (NLDN) stroke data. The Lapierre et al. analysis was conducted over three regions of the U.S. (Southeast, South Central, and North Central states) for May–August 2014–2015. To minimize contamination by pollution, all columns $>3 \times 10^{15}\text{ mol cm}^{-2}$ were removed before the analysis. They obtained a mean LNO₂ PE of 1.6 mol/stroke , which translates to $\sim 24\text{ mol LNO}_x\text{ fl}^{-1}$. They also found that the PE for CG strokes was $\sim 10\times$ larger than the PE for IC strokes and attributed a portion of the difference to differences in peak current and stroke rate. Q. Zhu et al. (2019) compared WRF-Chem computed flash rates using two combinations of convective and lightning parameterizations: Kain-Fritsch convection with a CAPE-Convective Precipitation flash rate scheme and Grell 3D convection with a cloud-top height flash scheme. Predicted lightning density over the southeastern US was better predicted with the former set of schemes. The resulting NO₂ AMF was 16% less using this set of schemes than using Grell/CTH in the southeast US in spring/early summer, which would result in larger BEHR vertical column NO₂ values.

The TROPOMI instrument flying on the Sentinel 5P satellite provided NO₂ columns initially at $3.5 \times 7\text{ km}$ resolution (much finer than OMI), and later at $3.5 \times 5.5\text{ km}$. Allen, Pickering, Bucsela, et al. (2021) were the first to use TROPOMI data for estimation of LNO_x PE. Analyzing 29 case studies over the CONUS and adjacent Atlantic, they obtained PE values of $175 \pm 100\text{ mol fl}^{-1}$ with GLM optical flashes and $120 \pm 65\text{ mol fl}^{-1}$ with ENTLN flashes. The uncertainty range considers the sensitivity of the PE to the a priori profile (GMI-based or cloud-scale modeling based (Ott et al., 2010)) used in the AMF and to the method of determining the background (10th or 30th percentiles of the retrieved vertical column amount).

Pérez-Invernón, Huntrieser, Erbertseder, et al. (2022) performed LNO_x analyses using TROPOMI data focused on the region of the Pyrenees and Ebro Valley in Spain. LNO_x was estimated from both the KNMI (van Geffen et al., 2022) and DLR TROPOMI retrievals. Flashes from the Earth Networks Global Lightning Network (ENGLN) and European Consortium for Lightning Detection (EUCLID; Schulz et al., 2016) networks were used. A priori profiles used in AMF calculations came from the ECHAM-MESSy Atmospheric Chemistry Model. Estimates of background were computed from non-flashing convective pixels. LNO_x PE values obtained were $47 \pm 33\text{ mol fl}^{-1}$ for the KNMI retrieval with ENGLN flashes; $69 \pm 34\text{ mol fl}^{-1}$ for the KNMI retrieval with EUCLID flashes; $58 \pm 22\text{ mol fl}^{-1}$ for the DLR retrieval with ENGLN flashes; and $51 \pm 25\text{ mol fl}^{-1}$ for the DLR retrieval with EUCLID flashes. Uncertainty due to the background estimate was judged to be larger than due to the choice of lightning data set. Pérez-Invernón, Gordillo-Vázquez, van der Velde, et al. (2023) analyzed 16 storm cases from the Ebro Valley, Spain using TROPOMI NO₂ and flash data from an LMA. A priori profiles came from the same model as in the previous analysis, but in this work the 10th and 30th percentile column amounts were assumed as background. For these 16 cases (Figure 10), a correlation of 0.77 was found between LNO_x PE and flash channel length. An exponential function best fit the relationship between PE and mean flash frequency, which is similar to the finding of Bucsela et al. (2019). Mean PE over the 16 cases was $108 \pm 82\text{ mol fl}^{-1}$ when the 10th percentile value was assumed for background and $54 \pm 42\text{ mol fl}^{-1}$ when the 30th percentile value was used.

Zhang et al. (2023) took advantage of the fact that TROPOMI orbital swaths overlap at high latitudes to obtain LNO_x observations in the Arctic (north of 70°) on successive orbits (100 min apart). Both the LNO₂ lifetime and PE can be obtained from these data. An assumed LNO₂ profile was used as the a priori in the AMF calculations

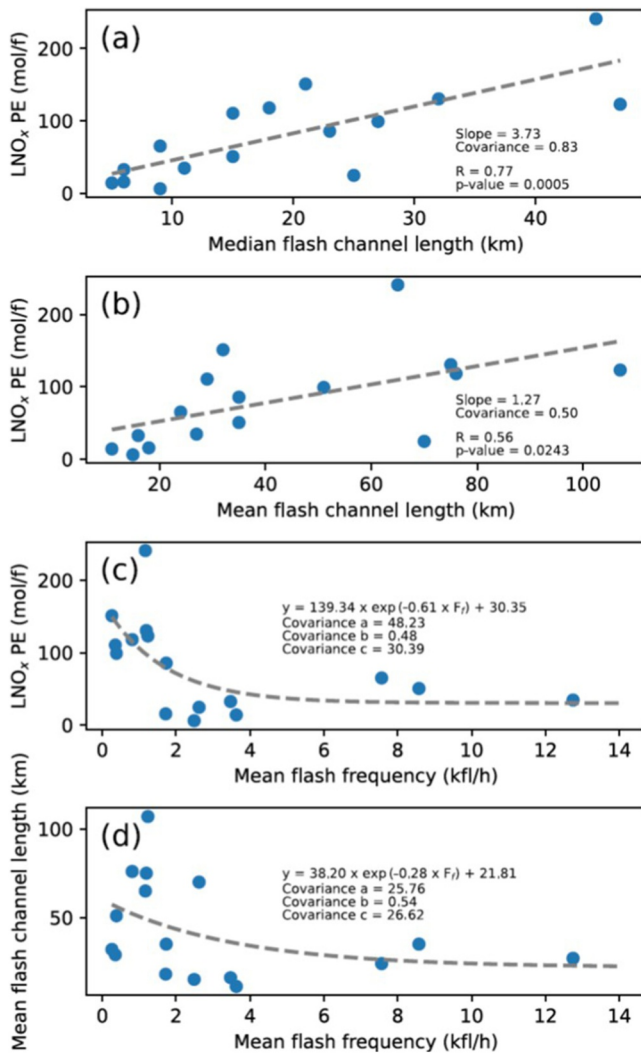


Figure 10. Scatterplots showing the relationship between LNO_x PE, flash channel length, and flash frequency. Panels (a) and (b) show the PE versus the median and mean flash channel length. Panel (c) shows the PE versus the mean flash frequency. Finally, panel (d) shows the mean flash channel length versus mean flash frequency. Sixteen cases were used to obtain the relationships and the linear regression (dashed line). The Pearson correlation coefficient (R) and the corresponding p -values are shown in panels (a), (b), as well as a fit to an exponential decay in panels (c), (d). From Pérez-Invernón, Gordillo-Vázquez, van der Velde, et al. (2023).

along with a 30th percentile background. Using the lightning data from the Vaisala Global Lightning Data set (GLD360) for the summer seasons of 2019–2021, it was found that LNO_2 PE was a factor of six larger over the ocean than over surrounding continents, consistent with the findings of Bucsela et al. (2019) and Allen et al. (2019). Higher LNO_x PE was often found with lower flash rates. Median LNO_2 PE was $0.9 \text{ mol stroke}^{-1}$ over land, $6.4 \text{ mol stroke}^{-1}$ over ocean and $6.5 \text{ mol stroke}^{-1}$ over coastal areas, with a range of at least a factor of 20 in each region. Unfortunately, Zhang et al. (2023) did not estimate PE per flash to allow comparison with other literature. However, the low values of PE per stroke do suggest that the multiplicity of flashes in this region was large. The LNO_2 lifetime in the near convection environment was estimated to be $6 \pm 2 \text{ hr}$ in the relatively clean Arctic environment, which is twice as long as estimated by Nault et al. (2016) for a midlatitude storm containing pollution. NO_x is lost more rapidly to organic nitrates in a polluted environment.

Even finer spatial resolution than TROPOMI has been achieved with the GeoCAPE Airborne Simulator (GCAS; Kowalewski & Janz, 2014). NO_2 columns from GCAS flown on the NASA ER-2 aircraft are available at 250 m resolution. Allen, Pickering, Lamsal, et al. (2021) used the GCAS NO_2 retrievals in LNO_x PE estimates for 10 convective systems over the US and the Western Atlantic during the GOES-R Post Launch Test Campaign in April and May 2017. GMI profiles were used in the AMF calculation, and the 5th percentile of the column amounts was assumed as the background. The high spatial resolution allowed identification of a small number of columns minimally affected by lightning. Separate PE estimates were made using data from the Geostationary Lightning Mapper instruments on the GOES satellites and the ground-based ENTLN. Mean PE over the 10 storms was $360 \pm 180 \text{ mol fl}^{-1}$ for GLM and $230 \pm 115 \text{ mol fl}^{-1}$ for ENTLN. PE was found to be positively correlated with GLM multiplicity and flash optical energy and negatively correlated with flash density (Figure 11). The strong relationship between PE and flash optical energy agrees with Wu et al. (2023). Storms with high flash rates (leading to high flash density) tend to have shorter channel length flashes (e.g., Bruning & MacGorman, 2013; Bruning & Thomas, 2015), and, therefore, smaller LNO_x PE per flash.

3.5. LNO_x in Regional and Global Models

The NO_x source due to lightning (LNO_x) is needed in atmospheric composition models, as this source is an important contributor to the reactive nitrogen budget. Neglecting this source leads to underestimation of NO_y and O_3 in the middle and upper troposphere. Models require parameterizations for estimating flash rates, estimates of the LNO_x PE for CG and IC flashes, and a method of distributing the LNO_x production in the vertical. Single parameter

flash rate schemes and multi-parameter schemes for use in regional or global models are described in Sections 3.5.1.1 and 3.5.1.2, respectively. Section 3.5.1.3 presents studies comparing the results of these parameterizations, and Section 3.5.1.4 describes how flash rates in models can be constrained by observations. Section 3.5.2 focuses on specification of LNO_x PE and its vertical distribution in models. This section also presents a summary of studies that used aircraft or satellite data to constrain LNO_x PE in global or regional models.

The lightning source (g_N) for a model grid cell and time step can be expressed as

$$\text{LNO}_x = [f_{\text{CG}} \times \text{LF} \times E_{\text{CG}} + (1 - f_{\text{CG}}) \times \text{LF} \times E_{\text{IC}}] \times P \times \text{CONV} \times \Delta t, \quad (2)$$

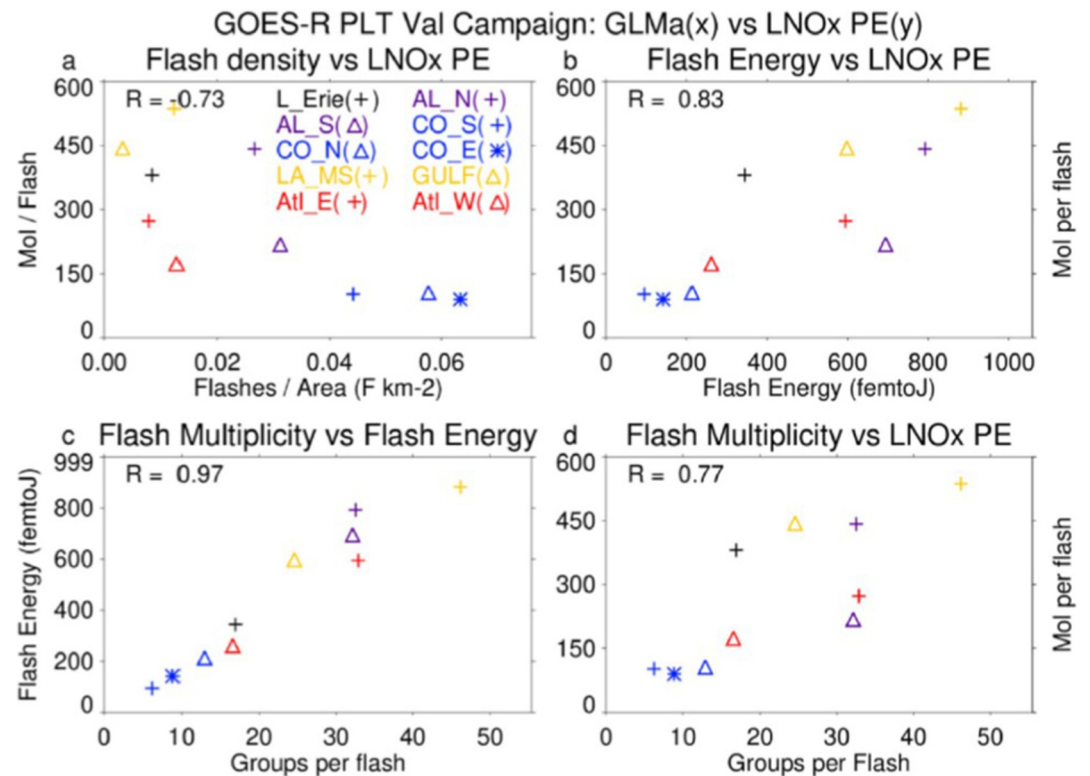


Figure 11. Scatterplots showing the relationship between (a) LNO_x PE (moles per flash) and flash density (flashes km^{-2}), (b) LNO_x PE and flash energy (fJ), (c) flash energy and flash multiplicity, and (d) LNO_x PE and flash multiplicity. Colors are used to separate flight days while symbols are used to separate system within each flight day. Correlations are shown in the upper left of each panel. From Allen, Pickering, Lamsal, et al. (2021).

where f_{CG} is the fraction of flashes classified as cloud-to-ground (CG), LF is the lightning frequency (flashes s^{-1}), E_{CG} and E_{IC} are the mean energies of CG and intracloud/intercloud (IC) flashes in joules (J), P is the mean NO production per unit energy (molecules NO/J), and CONV is a conversion factor equal to the molecular weight of nitrogen (14 g/mol) divided by Avogadro's number, and Δt is the time step in seconds.

The cloud-to-ground fraction (f_{CG}) is needed unless E_{CG} is assumed to equal E_{IC} . Price and Rind (1993) parameterize this fraction in terms of cold-cloud depth using a 4th order polynomial fit to observations of western U.S. thunderstorms. Whaley et al. (2024) assume f_{CG} increases with absolute latitude varying from 10% at the equator, 40% at $\pm 45^\circ$ to 70% at the poles based on observations of the freezing height in the clouds and resulting cloud-to-ground fraction (Uman, 1986). The IC:CG ratio has also been found to increase with storm intensity. Rutledge et al. (1992) found that the ratio over Darwin was proportional to the square root of the total flash rate, while Ren et al. (2024) found a power law relationship over three of four study regions in China.

3.5.1. Parameterization of Lightning Frequency (LF)

3.5.1.1. Single Parameter Schemes

Many global and regional atmospheric composition models parameterize the LF in terms of model fields such as those listed in Table 5. Most common is the use of the cloud-top height (Price & Rind, 1992) to predict flash rate. Slightly more sophisticated schemes parameterize the LF in terms of microphysical variables such as the downward flux of graupel, the upward flux of cloud ice (Finney et al., 2014), and/or the ice water path (Deierling et al., 2008; W. A. Petersen et al., 2005). The performance of several early flash rate schemes is compared in Tost et al. (2007).

With few exceptions, flash rates associated with the same values of predictor variables are greater over continental locations than marine locations. Therefore, separate continental and marine fits to data are useful to capture the

Table 5
Single Parameter Flash Rate Parameterization Schemes

Flash rate parameterization scheme	References
CAPE	Rutledge et al. (1992)
Cloud-top Height	Luhar et al. (2021); Michalon et al. (1999); Price and Rind (1992)
Cold Cloud Depth	Futyan and Del Genio (2007); Yoshida et al. (2009)
Convective Precip Rate	Allen and Pickering (2002); Allen et al. (2012); Huijnen et al. (2010); Kang, Pickering, et al. (2019); Meijer et al. (2001)
UT Convective MassFlux	Grewe et al. (2001); Allen and Pickering (2002)
Integrated ConvectiveMass Flux	Deierling et al. (2008)
Updraft vertical velocity or Volume	Deierling and Petersen (2008); Barthe et al. (2010)
Upward Flux of Cloud Ice	Finney et al. (2014); Romps (2019)
Ice Water Path	Deierling et al. (2008); Petersen et al. (2005)

large contrast between continental and marine flash rates. Separate continental and marine fits are preferred over land and water schemes because convection that is continental in its character may extend well offshore.

The cloud-height class of schemes (Luhar et al., 2021; Michalon et al., 1999; Price & Rind, 1992 (PR92)) parameterize flash rates in terms of powers of convective cloud top height, a measurable variable available in all standard output streams. The most popular scheme (PR92) assumes a 4.9th (1.7th) power dependence at continental (marine) locations. For example, each of the Earth-system models with interactive tropospheric chemistry participating in CMIP-6 (Eyring et al., 2016), parameterized flash rates in terms of PR92. These models consist of UKESM1 (Archibald et al., 2019), GFDL-ESM4 (Horowitz et al., 2020), CESM2-WACCM (Danabasoglu et al., 2020; Emmons et al., 2020; Lamarque et al., 2012), and GISS-E2-1 (Kelley et al., 2020). Other global models that use PR92 to parameterize flash rates include GEOS-CHEM (Murray et al., 2012), MOZART-4 (Emmons et al., 2010), and CHASER (Sudo et al., 2002). The power dependence of this robust class of schemes is supported by theory at least at continental locations (Boccippio, 2002), but the scheme does not incorporate microphysical information and is sensitive to small biases in the height of model convection. It is also known to overestimate the fraction of grid boxes with flashes and the fraction of flashes that occur in the tropics. Cloud-top height in global and regional models has been defined by a variety of methods: top of upper-most model layer containing hydrometeors, top of upper-most layer with non-zero convective mass flux, and level of neutral buoyancy.

Several alternatives to the PR92 cloud top height scheme also exist in the literature. Allen and Pickering (2002) parameterized flash rates in terms of upper tropospheric convective mass flux. Their approach was used in the GMI (Allen et al., 2010) and CMAM (Scinocca et al., 2008) models. Convective mass flux is also included in the Grewe et al. (2001) scheme. A downside of this approach is that observations of convective mass flux (needed for model validation) are difficult and the flash rates produced are highly dependent on the convective scheme. Flash rates in the global TM5 (Huijnen et al., 2010) and regional CMAQ (Allen et al., 2012) models are specified in terms of convective precipitation. A downside of this approach is that regional variations in the precipitation yield per flash are often large.

Finney et al. (2014) parameterized flash density in terms of upward cloud ice flux at 440 hPa, which they obtained by dividing the product of the 440 hPa IWC and 440 hPa convective mass flux by fractional cloud cover. They chose 440 hPa because it is located within the mixed phase region of a cloud and a cloud top pressure of less than 440 hPa is one of the criteria to be classified as a deep convective cloud according to the International Cloud Climatology Project (Rossow et al., 1996). Using separate fits over land and ocean, they found that their flash distribution more closely captured the observed spatial distribution and interannual peak-to-peak differences than existing schemes. For future climate simulations, Finney et al. (2018) adjusted the sampling level to 390 hPa ($\sim -40^{\circ}\text{C}$) to account for climate-induced changes in the top of mixed-phase cloud region. Romps (2019) also implemented an ice flux base scheme noting that the flux is related in a physical manner to the non-inductive charging mechanism. Romps (2019) chose to use ice flux values at 260 K instead of 440 hPa, because it is

Table 6
Multi-Variable Flash Rate Prediction Schemes

Reference	Variables	Model
Cloud resolving models		
McCaul et al. (2009)	Graupel flux; vertically integrated ice	HRRR
Yair et al. (2010)	For Lightning Potential Index (LPI): vertical velocity, cloud ice, graupel, snow	WRF
Lynn et al. (2012)	LPI, mass of ice	WRF
Global and regional models		
Romps et al. (2014); Tippett and Koshak (2018)	CAPE, total precipitation	GEFS
Liaskos et al. (2015)	Cold cloud depth, total cloud depth, CAPE, integrated cold-core convective mass flux, surface temperature	GEOS
Lopez (2016)	CAPE, vertical profile of frozen precip mass flux, profile of condensate in updraft, convective cloud base height	ECMWF GEOS
Stolz et al. (2017)	Normalized CAPE, CCN concentration, warm cloud depth	—
Etten-Bohm et al. (2021)	Normalized CAPE, LCL, column saturation fraction	CAM5
He et al. (2022)	CAPE, vertically integrated ice	CHASER
Long continuing current flashes (LCC)		
Pérez-Invernón et al. (2022b)	Updraft mass flux, ratio of LCC to total flashes	MESSy

believed that the charge separation zone and mixed phase region will continue to follow the 260 K isotherm even in a warming climate.

Using meteorological fields from WRF and flash counts from the NLDN, Kang, Pickering, et al. (2019) developed lightning parameterizations for use in the regional atmospheric chemistry model, CMAQ. For periods without hourly lightning data, they derived regression equations for 36 km² regions between modeled convective precipitation and NLDN flash rates using fields for 2002–2014. For retrospective simulations of periods and regions with lightning data, they suggested using flash rates derived from hourly NLDN observations. To provide consistency with meteorological fields, this approach is best used with assimilation of lightning data to ensure accurate location and timing of deep convection.

3.5.1.2. Multi-Parameter Schemes

Several flash rate parameterizations have been developed that use multiple variables as predictors. These schemes are summarized in Table 6. The first three schemes are best used in cloud-resolving models, whereas the remaining schemes are designed for regional or global models. We choose to include these three cloud-resolving schemes because they are used in operational forecasting models. More comprehensive reviews and tests of parameterization schemes for cloud-resolving models are presented by Bovalo et al. (2019) and Cummings et al. (2024). The WRF cloud-resolving model with two-moment microphysics can be run with an explicit electrical charging scheme and discharges that are allowed to occur when an electric field breakdown potential is reached (WRF-ELEC; Fierro et al., 2013).

McCaul et al. (2009) developed lightning prediction schemes based on the flux of graupel (updraft vertical velocity $\times$ graupel mixing ratio) and vertically integrated ice (graupel, snow, cloud ice). A weighted summation of these two methods provided the best prediction of observed flash rates. This scheme is used in the High-Resolution Rapid Refresh model run by the U. S. National Weather Service, as well as in convection-permitting simulations with the UK Met Office Unified Model (Field et al., 2018; Finney et al., 2020). Yair et al. (2010) used the lightning potential index (LPI) to predict the probability of lightning in the regional Weather Research and Forecasting (WRF) model. They obtained the LPI by integrating the product of the kinetic energy and the potential for charge separation over the main charging region of the cloud (0 to -20°C). The kinetic energy was determined by taking the square of the grid-scale vertical velocity. The potential for charge separation is a dimensionless quantity whose value depends on the concentrations of ice, graupel, snow, and liquid water. LPI was further developed by Lynn et al. (2012) into a lightning prediction algorithm, which considers the potential electric energy available for lightning production.

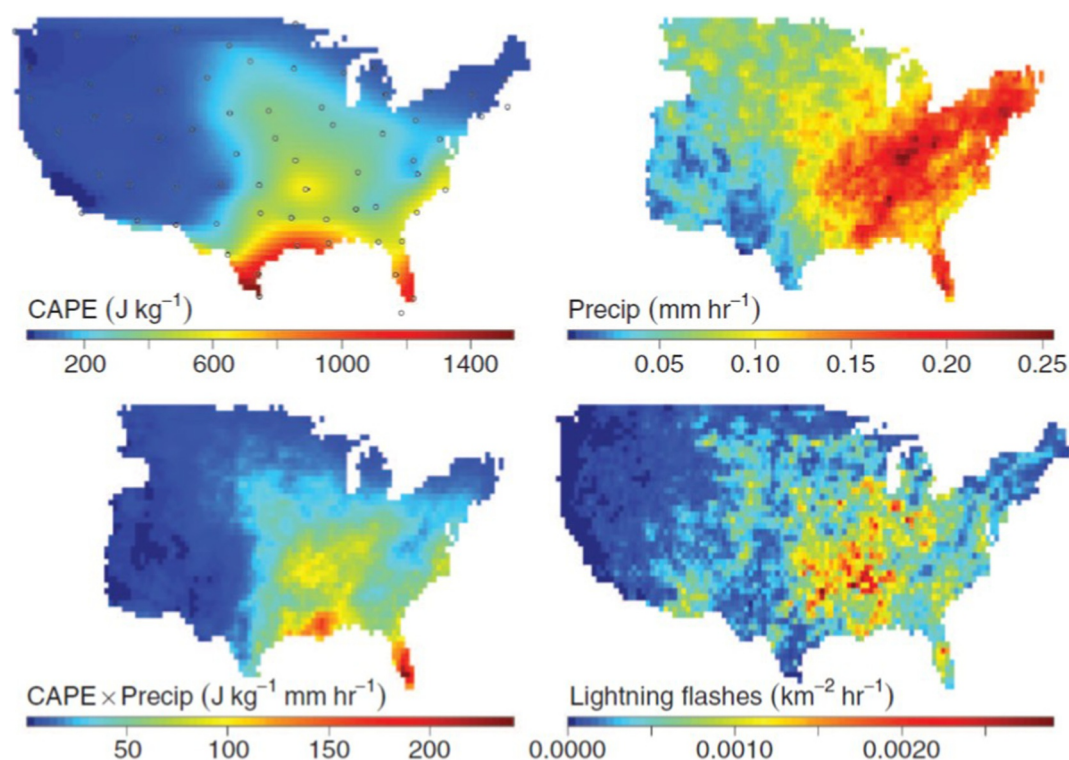


Figure 12. Mean maps of CAPE (top left), precipitation (top right), CAPE times precipitation (bottom left), and lightning flashes (bottom right) for the year 2011. CAPE was obtained from the SPARC radiosonde data, precipitation was obtained from the National Weather Service River Forecast Center, and CG lightning was obtained from NLDN. From Romps et al. (2014).

Romps et al. (2014) developed a parameterization that used the product of two variables (CAPE and precipitation; see Figure 12) to predict CG flash rate, finding a 12% increase in CG flashes per degree of warming. Tippett and Koshak (2018) used GEFS model CAPE and precipitation output to predict CG lightning flashes over CONUS. Forecasts of the daily numbers of CG flashes had statistical significance out to a 15-day lead time for both CONUS and for some subregions.

Liaskos et al. (2015) developed a flash rate parameterization for the GEOS-5 global Chemistry and Climate Model by regressing meteorological fields from a 5-year GEOS-5 GCM simulation with climatological flash rates from the spaceborne Optical Transient Detector and Lightning Imaging Sensors (OTD/LIS) (Cecil et al., 2014). The scheme used a weighted sum of five normalized terms derived from cold-cloud depth, total cloud depth, CAPE, integrated cold-core convective mass flux, and surface temperature. They obtained a reasonable flash rate distribution but overestimated Amazon Basin flash rates during March–May while underestimating central African flash rates.

Lopez (2016) proposed a flash rate prediction scheme that uses quantities diagnosed from the ECMWF convection scheme: CAPE, Q_R , and the convective cloud-base height (CCL), where Q_R is an estimate of the charging rate due to collisions between graupel and other hydrometeors in the mixed phase region. Q_R is calculated by integrating the product of the graupel mixing ratio and the sum of the mixing ratios of cumulus cloud liquid water and snow over the mixed phase region of the cloud. A decade long model experiment showed reasonable comparison with the OTD/LIS climatology on annual (Figure 13), seasonal and continental scales. Flash densities were found to be almost independent of model horizontal resolution.

Stolz et al. (2015) analyzed convective features from the Tropical Rainfall Measuring Mission Precipitation Radar (TRMM PR) and LIS, along with CCN concentrations from the GEOS-Chem model. They found that greater normalized CAPE (NCAPE; CAPE divided by the depth of the buoyancy layer) and greater CCN concentration led to greater flash density, and that warm cloud depth (WCD) modulated the NCAPE-CCN interaction with greater flash density with decreasing WCD. Stolz et al. (2017) developed a logarithmic multiple-linear

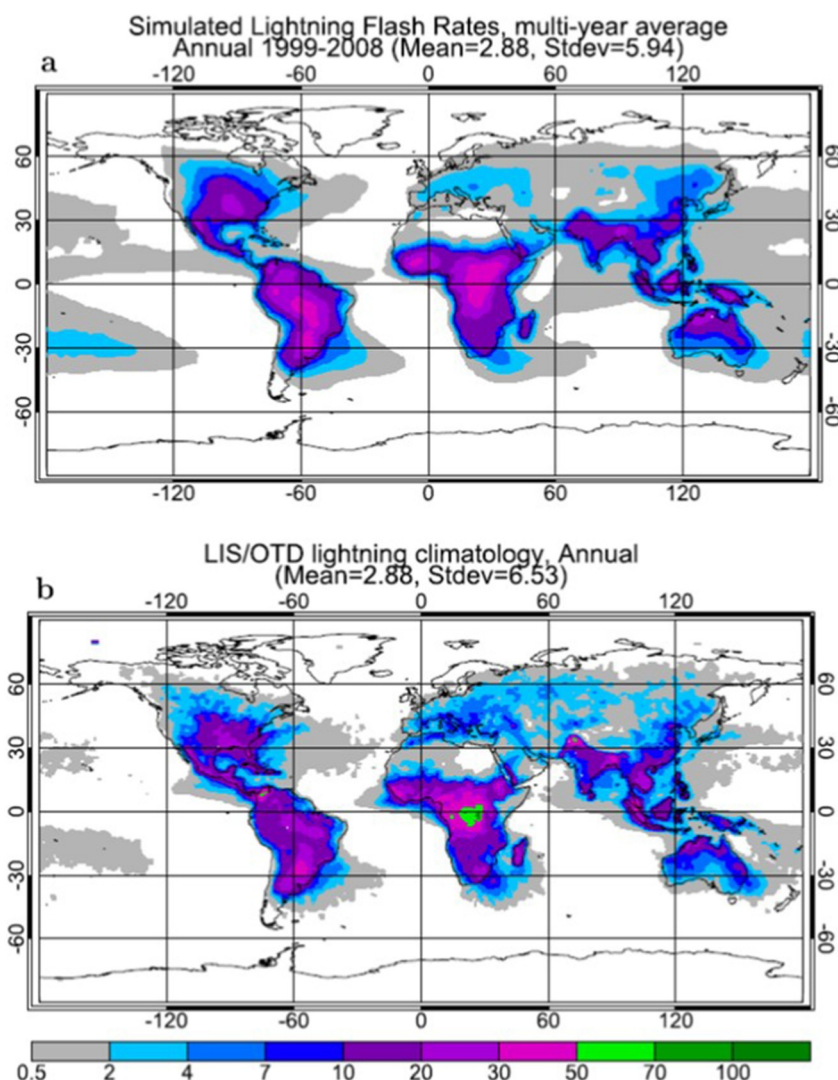


Figure 13. Annual mean lightning flash densities (a) from the 1999–2008 years-long 80-km global ECMWF model simulations and (b) from the LIS/OTD 1995–2010 climatology. Flash densities are expressed in flashes $\text{km}^{-2} \text{yr}^{-1}$. Global mean and standard deviation are reported at the top of each panel. From Lopez (2016). Copyright American Meteorological Society. Used with permission.

regression equation for flash density through the analysis of these variables. The importance of the aerosol effect on flash rates has been examined by several investigators, especially over the Amazon Basin (Albrecht et al., 2011; Allen et al., 2024; Altaratz et al., 2010, 2017; Wall et al., 2014). The Stolz et al. (2017) analysis plus the Amazonian evidence suggest that further work on incorporation of aerosol in lightning prediction algorithms should be conducted.

Etten-Bohm et al. (2021) developed the flash rate scheme implemented in version 5.1 of the Canadian Earth System Model (CanESM5.1) and CAM5. In this scheme, the probability of lightning occurrence is determined at each grid box and then multiplied by a global constant chosen to ensure that the flash rate within the OTD/LIS coverage region matches that of OTD/LIS. The probability of lightning occurrence was specified with a six-parameter fit to normalized values of CAPE, lifting condensation level (LCL), column saturation fraction (f_{sat}), a measure of how humid a column is relative to its saturation specific humidity, and their first order products ($\text{CAPE} \times \text{LCL}$, $\text{CAPE} \times f_{\text{sat}}$, and $\text{LCL} \times f_{\text{sat}}$). Whaley et al. (2024) evaluated this scheme in CanESM5. After removing the flash rate dependency on LCL over mountainous regions, they found that the model land-ocean flash-ratio value of four compared well with the observed ISS-LIS ratio value of five. They

noted that the ratio of four was much higher and closer to observations than that obtained with several other schemes that do not use different approaches over land and water or include a variable that effectively accounts for the different environments (e.g., Charn & Parishani, 2021).

Pérez-Invernón, Huntrieser, Jöckel, and Gordillo-Vázquez (2022) developed a method to parameterize long-continuing current (LCC) flashes (flashes with durations >9 ms) in the Modular Earth Submodel System (MESSy) using existing parametrizations of flash rates and a quadratic relationship they observed between the updraft mass flux and the ratio of LCC flashes to total flashes. A LCC parameterization is useful because LCC flashes are more likely to ignite wildfires or trigger sprite discharges in the mesosphere than flashes in general. They found the best agreement between modeled and observed LCC flashes was obtained when total flashes were parameterized using a cloud-height based scheme over land and a convective precipitation-based scheme over water.

One method to avoid separate schemes for continental and marine locations is to include variables in the scheme that vary greatly between continental and marine environments. Etten-Bohm et al. (2021) avoided separate schemes by including the lifting condensation level (LCL) as a term in their parameterization; LCLs and consequently cloud bases are typically lower over moist areas, such as the ocean (Williams & Stanfill, 2002). Lopez (2016) avoided separate schemes by including the convective cloud base height (CCL) in his parameterization, although technically the formula he uses to diagnose the concentrations of hydrometeors includes a dimensionless constant that varies between land and water locations.

As the factors contributing to lightning production are complex and difficult to constrain using physical formulations that consider only a few variables, machine learning approaches are increasingly being introduced to predict the occurrence and/or the density of flashes.

N. Liu et al. (2022) used the Convective Feature data base from TRMM that spans 1998–2023, meteorological variables (CAPE, CIN, low-level wind shear, and warm cloud depth (WCD)) from the ERA-Interim Reanalysis, and random forest models to predict the percentage of convective features that have lightning across the TRMM domain. The geographical distributions of the predicted and observed features were excellent with a spatial correlation coefficient of 0.88 and a perhaps surprisingly well-reproduced land-sea contrast. However, the models struggled to predict the timing and location of individual events with relatively low global skill scores (0.17 for Critical Success Index (CSI) and 0.26 for Heidke Skill Score (HSS)).

Ehrensperger et al. (2025) utilized a deep neural network trained on $0.25^\circ \times 0.25^\circ$ ERA5 tropospheric profiles of nine meteorological variables with summertime 2010–2019 lightning observations (flash or no-flash) over the Eastern Alps from the Austrian Lightning Information & Detection System (ALDIS). Using explainable artificial intelligence techniques (XAI), they determined that upper tropospheric cloud ice and mid-tropospheric snow water content (graupel) contributed the most to the model's determination of flash or no-flash. They noted that being able to train a neural network directly on atmospheric soundings coupled with XAI is a powerful tool for understanding the processes contributing to flashes and the causes of stark land-sea contrasts in flash rates.

Convolutional neural networks (CNNs) are a promising tool for lightning prediction because their two-dimensional input fields provide the network with information on the synoptic pattern that is not available from a local approach. Kalashnikov et al. (2024) used a CNN to predict the daily occurrence of cloud-to-ground lightning over the western U.S. from the spatial pattern of seven meteorological variables. The locally trained CNNs were skillful with a median precision-recall area under the curve (AUC) of 0.8, performing best in regions with plentiful lightning such as the interior Southwest. Feature analysis revealed that ratio of surface moist static energy (MSE) to free tropospheric saturation MSE, the lower troposphere lapse rate, and the RH played substantial roles in predicting flash occurrence.

P. Cheng et al. (2024) and W.-Y. Cheng et al. (2024) trained three supervised ML algorithms (neural network, random forest, and CNN) over the CONUS using seven standardized meteorological variables from ERA5, standardized precipitation data from TRMM, and flash density data from WWLLN. Comparisons of the ML-predicted flash rates with those from Roms et al. (2014) showed a reduction in RMSE over the southeastern U.S. where flash rates are high, especially when the non-local CNN was used; however, RMSE increased over the western U.S. where flashes are less frequent. Feature analysis revealed that the inclusion of TRMM precipitation data was critical with a 30% decrease in model performance when TRMM data were excluded.

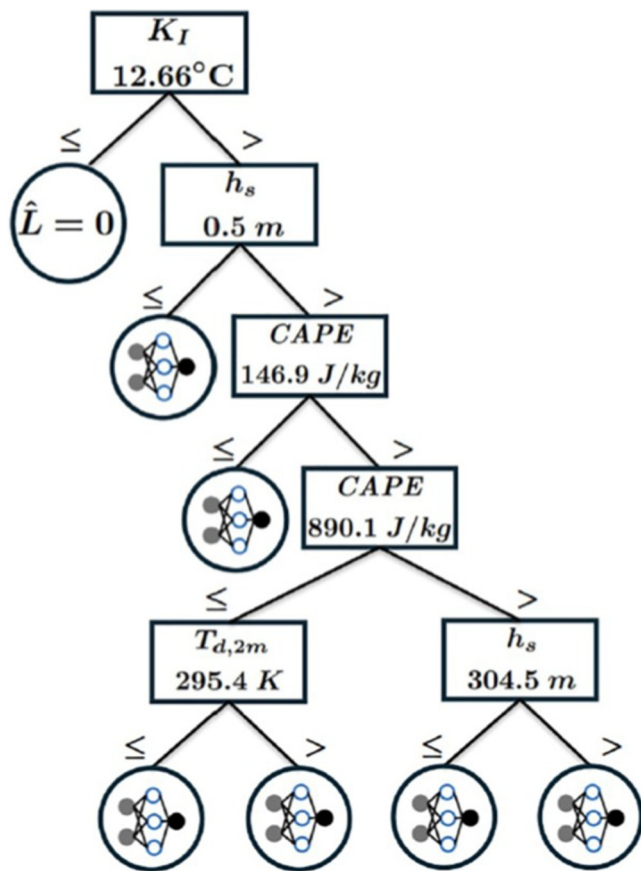


Figure 14. Schematic of the architecture of the calibrated ELM-tree as described in the text. Splitting values are given at each tree node. Inequality symbols show the direction of the splitting rules. The left split at the tree root corresponds to the zero-branch ($L = 0$) leaf node, which always returns 0 flashes. In the non-zero-branch, each leaf node consists of an Extreme Learning Machine (ELM) model, as illustrated with the sketches of single hidden layer feedforward neural networks. K_I = K Index. h_s = surface elevation. T_d = dew point at 2 m. From Verjans and Franzke (2025).

Using smoothed lightning data from Lightning Imaging Sensors (LISs) on TRMM and ISS and reanalysis data from ERA5, Verjans and Franzke (2025) used a decision tree approach to develop a flash rate scheme that they implemented in the Community Earth System Model version 2.2 (CESM2.2, Danabasoglu et al., 2020). After considering tradeoffs between complexity and accuracy, a tree of depth 4 (see Figure 14) was obtained. (a) A K-Index threshold was used to split the tree into a non-flashing node and a flashing branch. (b) The flashing-branch was then split into a maritime node and a continental branch using a surface elevation threshold. (c) The continental branch was then split into two nodes and a branch using values of CAPE. (d) The high-CAPE branch was then split into four nodes using thresholds of 2-m dewpoint and surface elevation. Extreme Learning Models (ELMs) were then used to predict the flash rate for grid boxes in each node. The most common fit-features were found to be: CAPE, surface latent heat flux, absolute latitude, total totals index, cloud top height, surface sensible heat flux, 10-m wind speed, cloud base height, K-index, and surface pressure. The spatial correlation with LIS was impressive (0.92) and the contrast between Central Africa, the Amazon, and Maritime Continent flash rates was reasonable.

3.5.1.3. Comparison of Flash Rate Schemes

Luo et al. (2017) simulated flash rates over the United States during July 2010 using several common parameterization schemes and meteorological fields from WRF. They obtained the best agreement with NLDN flash counts when they applied a power relationship to the parameter pairs CAPE and cloud-top height. They obtained the poorest results when they parameterized flash rates in terms of precipitation ice mass and convective precipitation rate possibly due to deficiencies in the model microphysics simulation as opposed to deficiencies in the scheme itself.

Gordillo Vazquez et al. (2019) examined the performance of six lightning parameterizations in CAM5. They found that ice flux-based schemes and cloud top height schemes produced the most realistic spatial correlations.

He et al. (2022) implemented three additional lightning schemes into the CHASER (MIROC) chemistry-climate model and compared flash rates from those schemes with those from the existing cloud-top height based scheme:

- The first of the new schemes assumed flash rates were proportional to the cloud ice flux at 440 hPa (Finney et al., 2014) or 260 K (Romps, 2019).
- The second scheme was the Lopez (2016) parameterization that had been implemented in the ECMWF forecast system. It parameterized flashes in terms of Q_R , CAPE, and the convective cloud base (CCL) height.
- The third scheme was derived by modifying the ECMWF scheme to be consistent with McCaul et al. (2009). It used CAPE and vertically integrated ice.

He et al. (2022) found that the ECMWF-MCCAUL scheme performed best with a spatial correlation of 0.83 and a RMSE of $3.20 \text{ fl km}^{-2} \text{ yr}^{-1}$. The correlation coefficients for the ICEFLUX (Finney), ECMWF (Lopez), and cloud-top height schemes were 0.80, 0.79, and 0.78, respectively. Figure 15 illustrates the global flash distributions from OTD/LIS and these four parameterization schemes.

Stolz et al. (2021) compared the spatial distributions of flash rates obtained using the PR92, Finney et al. (2014), and Stolz et al. (2017) schemes. They found that flash density estimates from the more complex Finney et al. (2014) and Stolz et al. (2017) schemes were more consistent with observations than those from the simple PR92 scheme.

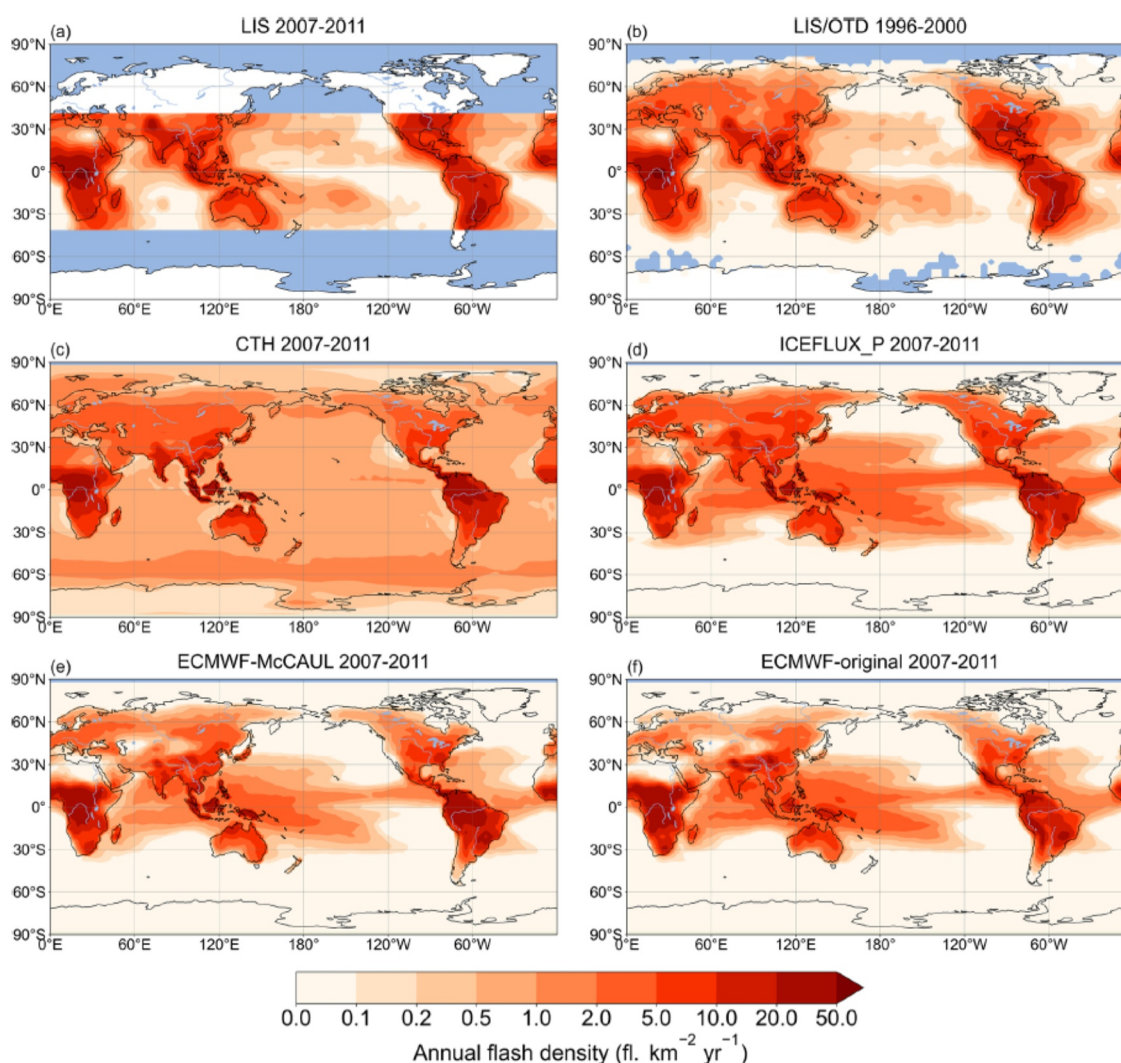


Figure 15. Annual mean lightning flash densities from (a) LIS satellite observations spanning 2007–2011, (b) LIS/OTD satellite observations spanning 1996–2000 but with a wider latitude range, (c) the CTH scheme in 2007–2011, (d) the ICEFLUX_P scheme in 2007–2011, (e) the ECMWF-McCAUL scheme in 2007–2011, and (f) the original ECMWF (Lopez, 2016) scheme in 2007–2011. From He et al. (2022).

3.5.1.4. Constraints on Flash Rate Schemes

To determine the subsequent O₃ production (see Section 5) accurately, LNO_x injection should be consistent in time and space with model-calculated deep convective mixing. Thus, most models continue to parameterize flash rates in terms of fields from the meteorological model used to drive the simulation even if continuous lightning observations are available. However, use of continuous flash data to place LNO_x may be a viable approach if coupled with assimilation of lightning data as the assimilation step will improve the location and timing of deep convection thereby minimizing inconsistencies with the meteorological fields (Heath et al., 2016; Kang et al., 2022; Vargas et al., 2024).

In some applications, interannual variability in model flash rates has been constrained using flash rates obtained from the NLDN, OTD/LIS, or Schumann resonances. For example, several investigators used flash rates from the National Lightning Detection Network (NLDN) and OTD/LIS to provide local/regional seasonal constraints on model flash rates (Allen et al., 2010, 2012; Kang, Pickering, et al., 2019; Murray et al., 2012). Murray et al. (2012) constrained cloud-height based flash rates in GEOS-Chem with regional scaling factors to ensure a monthly match with the observed LIS/OTD climatology. The regions were chosen using a hierarchical tree. After experimentation, they chose 37 regions because that choice resulted in well simulated monthly mean flash rates,

while avoiding excessively large instantaneous flash rates. X. Wang et al. (2025) constrained monthly flash rates in GEOS-Chem using observations of Schumann Resonance (SR). Global lightning activity can be derived from measurements of SRs, which are low-frequency electromagnetic waves within the atmosphere, primarily generated by lightning activity and resonating between the surface and the ionosphere. They found that the SR-constrained cloud-height based scheme did a better job of capturing interannual variations and decadal trends in global flash rates than the unconstrained scheme, which showed an unobserved 6.4% increase per decade in modeled lightning over the 2013–2021 time period.

A downside of constraints such as these is that they are only appropriate for simulations driven by assimilated meteorological fields and are likely inappropriate for simulations of the past or future with chemistry and climate models (CCMs).

The sensitivity of flash rate and LNO_x production to meteorology was examined (Zhao et al., 2009) by driving the regional scale model (REAM) with meteorological fields from MM5 and WRF. The WRF-REAM and MM5-REAM simulations were very similar when lightning NO_x was excluded. When including LNO_x , MM5-REAM simulated NO_2 columns were greater than WRF-REAM and satellite-retrieved SCIAMACHY columns likely due to an overestimation of the flash rate due to a high-bias in convective cloud top heights obtained from the Grell scheme used in MM5. The high-bias exists because the number of *intracloud* flashes in REAM depends on the cloud-top height; the CG flash rate is parameterized in terms of convective mass flux and CAPE. Comparison with INTEx-A measurements revealed that both simulations showed a larger than observed decrease in NO_x with latitude, likely due to latitude-dependent biases in convective mass flux.

Barret et al. (2010) investigated the impact of LNO_x and subsequent convective transport on upper tropospheric composition during the African Monsoon Multidisciplinary Analysis (AMMA) campaign by comparing results from simulations with four global CTMs to observations. The CTMs differed in their convective parameterizations, flash rate and LNO_x parameterizations, and total LNO_x source. The simulations confirmed that contribution of African LNO_x is most important over the tropical Atlantic. Comparison with satellite data, revealed that the convective cloud top heights in p-TOMCAT were too low. After reducing the entrainment/detrainment rates, mixing of stable environmental air into the cloud was lessened, allowing the storms to reach higher and more realistic altitudes improving convective transport, the flash rate and LNO_x source.

3.5.2. Parameterization of LNO_x PE

This section first discusses several issues concerning the representation of LNO_x in global and regional models, followed by presentation of LNO_x PE estimates obtained when simulations with these models are constrained by constituent data from either aircraft or satellite.

3.5.2.1. Factors Responsible for Flash-to-Flash Variations in LNO_x PE

The mean energy of lightning flashes varies with channel length, conductivity, and channel radius and is usually assumed to be larger for CG flashes than IC flashes. Price et al. (1997) assume that CG flashes have 10 times as much energy (E_{CG}) as IC flashes (E_{IC}) (see Lightning Energies subsection of Section 3 of Price et al.). E_{CG} is assumed to be approximately 3 times greater than E_{IC} in the UKCA StratTrop model following Schumann and Huntrieser (2007); however, in the CESM2 and GISS-E2.1 models, E_{CG} is assumed to equal E_{IC} as recommended by Ridley et al. (2005) and Ott et al. (2010) based on the combination of in-cloud NO_x observations and cloud-resolved modeling. In summary, the ratio of E_{CG} to E_{IC} still remains one of the largest uncertainties in parameterizing LNO_x in models.

Several studies indicate that the LNO_x production per flash depends on the flash channel length, which is inversely correlated with the flash frequency (e.g., Bruning & Thomas, 2015; Bucselá et al., 2019; see Section 3.4) and may also be longer in extratropical regions than in tropical regions (Huntrieser et al., 2009). Motivated by this relationship, Pérez-Invernón et al. (2025) implemented the flash-rate dependent PE formula from Bucselá et al. (2019) in the ECMWF–Hamburg (ECHAM)/Modular Earth Submodel System (MESSy version 2.55) Atmospheric Chemistry (EMAC) model (Jöckel et al., 2016) and compared the flash rate distribution from that scheme to that obtained using Price et al. (1997). The use of the power function relationship to flash frequency resulted in enhanced production of $\text{LNO}_x \text{ fl}^{-1}$ in low flash rate storms resulting in more $\text{LNO}_x \text{ fl}^{-1}$

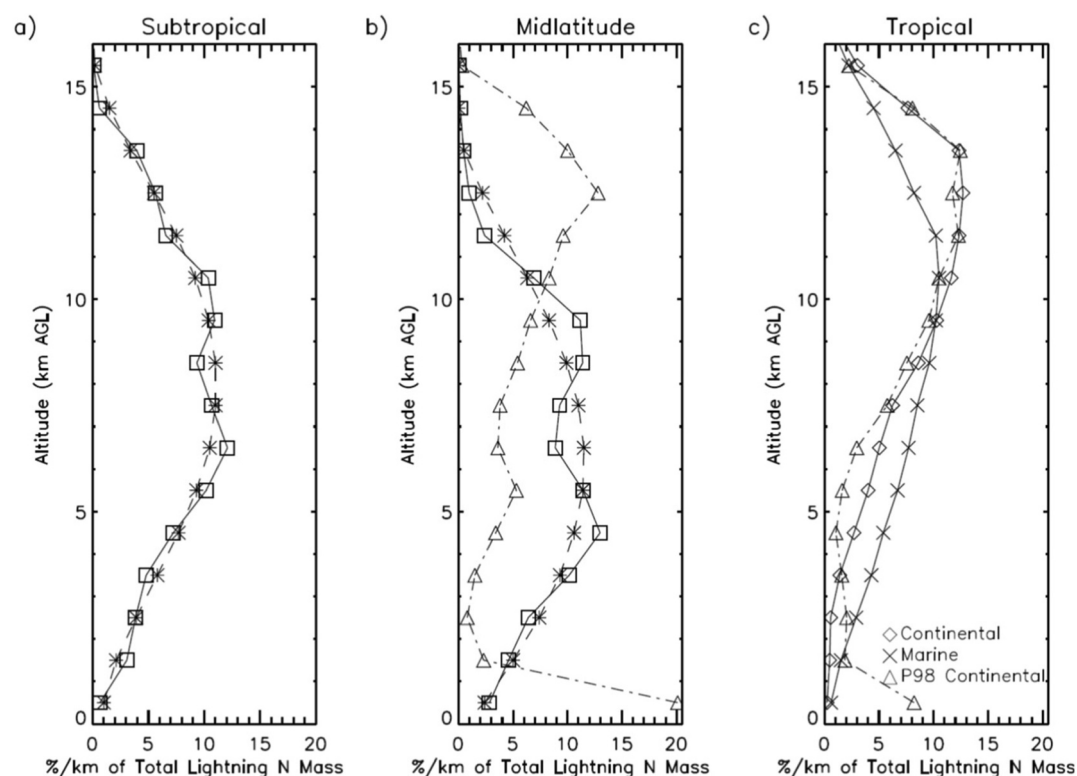


Figure 16. Average vertical distribution of percentage of LNO_x mass per kilometer following convection (solid) for the (a) subtropical and (b) midlatitude continental regimes. Dashed line shows polynomial fit. Midlatitude continental profile from Pickering et al. (1998) (dash-dot) is also shown in (b). The hypothetical tropical marine profile (c) was created by extrapolating the subtropical average profile to a higher tropopause regime while the tropical continental profile was constructed using the Pickering et al. (1998) profile with the boundary layer maximum removed and that mass redistributed into layers from 4 to 11 km. From Ott et al. (2010).

over water and during the winter season. The net result was a more spatially homogeneous distribution of LNO_x corresponding better to reality.

3.5.2.2. Vertical Placement of LNO_x

Lightning- NO_x is primarily released in the free troposphere with the method used to allocate emissions in the vertical varying from model to model. Such emissions have been distributed in a number of different ways:

- With a C-shaped profile that has a maximum in the upper troposphere with a smaller downdraft dependent peak in the lower troposphere (Pickering et al., 1998)
- Contrarily with a reversed-C-shaped profile that places the bulk of the emissions in the middle to upper troposphere (Ott et al., 2010)
- Linearly in log of pressure or pressure (Archibald et al., 2019)

Figure 16 illustrates the differences between the Pickering et al. (1998) and Ott et al. (2010) profiles.

These schemes incorporate both the vertical pattern of LNO_x emissions and vertical transport by updrafts and downdrafts. Alternatively, the LNO_x can be placed in the model based solely on the vertical distribution of emissions and allow the model convection to vertically redistribute the NO_x (e.g., Allen et al., 2012; DeCaria et al., 2005; Kang, Pickering, et al., 2019). Pre-convective vertical distributions of emissions have also been derived from the segment altitude distributions (SADs) of lightning channel segments observed by LMAs (Koshak, 2014; Koshak, Peterson, et al., 2014).

Luo et al. (2017) obtained pre-convection LNO_x profiles using SADs from the North Alabama Lightning Mapping Array (NALMA), flash counts from NLDN, and two different LNO_x parameterizations within LNOM. They then used the parameterizations in a simulation with WRF-CMAQ. In general, the post-convection NO_x

vertical distributions were very similar indicating that the choice of initial LNO_x profiles may be less important than the choice of parameterization scheme used to convectively redistribute LNO_x.

Mecikalski and Carey (2018) analyzed radar and NALMA data to develop 2-D histograms of IC, CG, and hybrid flash initiation and propagation locations relative to altitude, temperature, and radar reflectivity for multicell, supercell, and mesoscale convective systems. Differences between these profiles and those from the literature discussed above were noted, and it was recommended that these types of profiles be used in modeling LNO_x.

3.5.2.3. Observation-Based Constraints on LNO_x Production per Flash

Several studies have adjusted the NO_x yield per flash in global or regional models to best match trace gas profiles from field campaigns and/or trace gas columns from satellite-borne instruments. Early attempts using this method were reported by Schumann and Huntrieser (2007). More recent activity is reported below and is summarized in Table 7. If necessary, global total LNO_x production reported in these papers has been converted to PE in moles LNO_x per flash using a global climatological flash frequency of 46 fl/s (Cecil et al., 2014).

Using forward trajectories from observed lightning, Jourdain et al. (2010) determined that enhancements in Tropospheric Emission Spectrometer (TES) O₃ layers were related to O₃ production downwind of convective events. They found that GEOS-Chem when run with a PR-based scheme constrained regionally and monthly by OTD/LIS climatology did a good job of capturing the enhancements. GEOS-Chem, when run in this mode, captured the ozone enhancements seen by TES although the modeled signature was usually smaller and less variable than observed. They obtained an improved O₃ distribution when they increased the LNO_x source from 250 to 520 mol per flash but cautioned that the model bias could be due to factors other than the magnitude of the LNO_x source.

Parameterizing flash rates in terms of 440 hPa convective mass flux, Martini et al. (2011) examined the sensitivity of atmospheric composition to PE values of 240 and 480 mol per flash by comparing output from the University of Maryland Chemistry and Transport Model (UMD-CTM) to DC-8 measurements from the INTEX-A campaign (Singh et al., 2007). INTEX-A was conducted in the summer of 2004 over North America and the Atlantic. In the upper troposphere, they found that the larger source led to 30%–40% reduced biases for NO, NO_x, and OH, slightly higher biases for O₃ and PAN, but more than a factor of two higher biases for NO_y and HNO₃. Thus, due to the inconclusive results, they presented results for both source magnitudes in their paper.

Murray et al. (2012) implemented a cloud-height based scheme in GEOS-Chem that assumes a yield of 500 mol of N per flash (Hudman et al., 2007) for extratropical flashes north of 23°N in America and 35°N in Eurasia, and 260 mol N per flash for other flashes. The extratropical and tropical values for LNO_x PE were obtained by comparing modeled and observed NO₂ profiles and columns from ICARTT, SCIAMACHY, and the tropics.

M. Cooper et al. (2014) used GEOS-Chem with the Murray et al. (2012) flash rate parameterization to simulate atmospheric composition in 2008. They found that the standard simulation greatly underestimated column amounts of HNO₃ over southeast Asia. They found that the column bias could be eliminated by increasing the LNO_x over southeast Asia by a factor of 3.5 (to 2.4 Tg N yr^{−1}). However, that led to a large high bias in O₃. They hypothesized that the high-bias was caused by overly efficient O₃ production in GEOS-Chem due to unrealistically diluted LNO_x plumes. After implementing a sub-grid plume parameterization for LNO_x, they were able to obtain realistic HNO₃ and O₃ distributions over the region with a LNO_x source increase of just 0.5 Tg N yr^{−1}, as opposed to the 2.4 Tg increase needed without a sub-grid plume parameterization.

Miyazaki et al. (2014) were the first to conduct an extensive assimilation of atmospheric composition data from satellites. They parameterized flash rates in the CHASER model in terms of cloud top height (Price et al., 1997) using a global tuning factor that ensured that the annual global flash rate matched that of the OTD/LIS climatology. Then, they optimized LNO_x emissions through assimilation of OMI NO₂, TES and MLS O₃, MLS HNO₃, and MOPITT CO (see Figure 17). The optimization led to a 56% increase in the LNO_x source over water and a 32% increase over land with the total magnitude increasing from 4.7 to 6.3 Tg N yr^{−1} (231–310 mol per flash). By taking advantage of the varying vertical sensitivities of the assimilated fields, they were also able to constrain the vertical distribution of LNO_x. The peak in the updated profile was lowered by up to 1.2 km from that in the Pickering et al. (1998) profile, with the largest decreases over the tropical western Pacific.

Table 7
Summary of Use of Global and Regional Models in Estimating LNO_x PE

Reference	Model	Region	Flash parameterization	Chemical constraints	LNO _x PE (mol fl ⁻¹)
Jourdain et al. (2010)	GEOS-Chem	Global	Cloud-top height constrained regionally to match monthly OTD/LIS	TES O ₃ profile	520
Martini et al. (2011)	GEOS met. UMD-CTM	Global; CONUS analysis	Convective mass flux at 440 hPa; constrained locally to match monthly NLDN or OTD/LIS	DC-8 aircraft: NO, NO ₂ , NO _y , HNO ₃ , PAN, O ₃ , OH	240 based on NO _y , HNO ₃ 480 based on NO _x
Murray et al. (2012)	GEOS-Chem	Global	Cloud-top height constrained regionally to match monthly OTD/LIS	SHADOZ O ₃ profiles; OMI/MLS tropospheric O ₃	Northern midlatitudes: 500; elsewhere 260
Cooper et al. (2014)	GEOS-Chem	Global; SE Asia analysis	Same as Murray et al. (2012), but with subgrid LNO _x plumes	IASI HNO ₃ columns	Same as Murray et al. (2012) except increased PE in SE Asia
Miyazaki et al. (2014)	CHASER	Global	Cloud-top height; annual global flash rate constrained to OTD/LIS	Assimilated OMI NO ₂ , TES and MLS O ₃ , MLS HNO ₃ , MOPITT CO	310
Liaskos et al. (2015)	GEOS with GMI chemistry	Global	Convective mass flux at 440 hPa; constrained locally to match monthly OTD/LIS	OMI NO ₂ , TES O ₃ , OMI/MLS O ₃ , SHADOZ O ₃	246
Nault et al. (2017)	GEOS-Chem	Global	Cloud-top height constrained by monthly OTD/LIS	DC-8 NO ₂ , HNO ₃ OMI NO ₂	450
Marais et al. (2018)	GEOS-Chem	Global	Cloud-top height constrained by monthly OTD/LIS	OMI UT partial columns	280 for both tropics and midlatitudes
Kelley et al. (2020)	GISS-E2-1	Global	Cloud-top height constrained by land and marine OTD/LIS	CH ₄ lifetime	290
Horner et al. (2024)	GEOS-Chem	Global	Cloud-top height constrained by monthly OTD/LIS	TROPOMI UT partial columns	<500 in northern mid-latitudes
Laughner and Cohen (2017)	WRF-Chem	CONUS	Level of neutral buoyancy	DC-8 aircraft NO ₂	500–665
Mean ± 1 sigma (most likely range)					374 ± 117

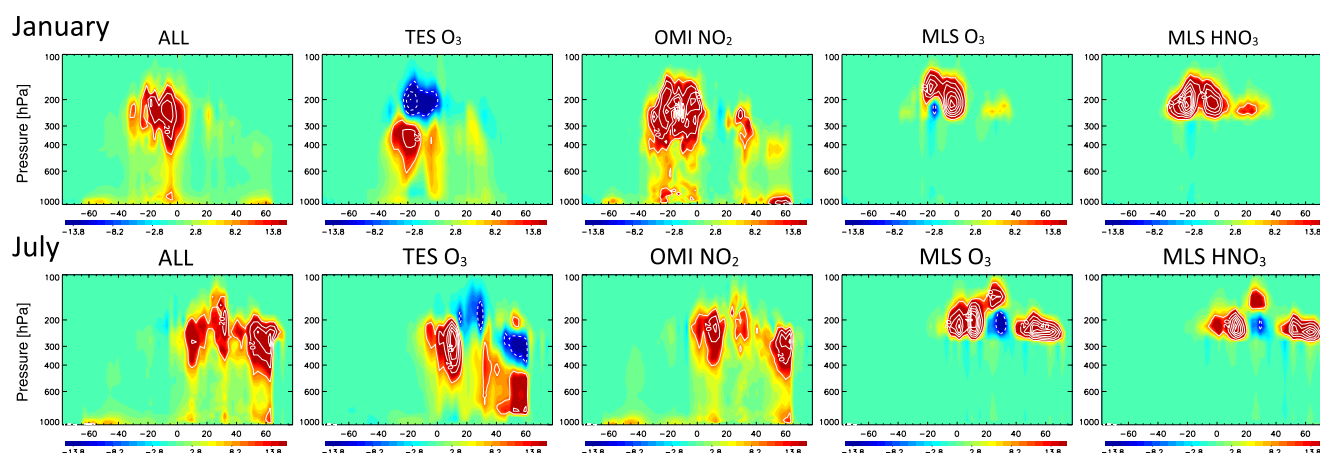


Figure 17. Latitude-pressure cross sections of the monthly mean analysis increment (the assimilation minus the CTM simulation) for the LNO_x source (in pptv day^{-1}) obtained from assimilation of (left) all the data, (2nd from left) TES O_3 data, (center) OMI NO_2 data, (2nd from right) MLS O_3 data, and (right) MLS HNO_3 data in (top) January 2007 and (bottom) July 2007. The interval of the contour lines is 10 pptv day^{-1} . From Miyazaki et al. (2014).

Liaskos et al. (2015) used the GEOS-5 model with GMI chemistry to examine the sensitivity of several trace gases to the magnitude of the LNO_x source for PE values ranging from 123 to 492 mol per flash. Focusing on September–December 2007, they found that increasing the PE by a factor of 4 led to tropical upper tropospheric enhancements of greater than 100% in NO_x , OH, HNO_3 , and PAN. This increase in LNO_x production led to O_3 enhancements of up to 60%, although the factor of 4 larger LNO_x led to significant positive bias in tropospheric O_3 . They found that an LNO_x source of 246 mol per flash agreed best with OMI NO_2 , TES O_3 , OMI/MLS O_3 , and SHADOZ O_3 profiles.

Laughner and Cohen (2017) examined the sensitivity of AMFs used to convert the slant column density (SCD) of NO_2 to a vertical column density (VCD) of NO_2 to the LNO_x source in the regional WRF-Chem model. They found that excluding LNO_x led to a 25%–35% low bias in the AMF. Through comparison with NO_2 profiles from the DC3 field campaign, they found that the optimal magnitude of the LNO_x source was relatively high, 500–665 mol per flash.

Nault et al. (2017) revisited estimates of LNO_x PE assuming a 3-hr lifetime for NO_x in the vicinity of deep convection due to loss of NO_x via reactions that produce methyl-peroxy nitrates ($\text{CH}_3\text{O}_2\text{NO}_2$), alkyl nitrates, and HNO_3 . They found that assuming a PE of ~ 450 mol per flash with a three-hour NO_x lifetime in the GEOS-Chem model agreed best with near field measurements of NO_2 and HNO_3 from the DC3 campaign and far field measurements of the same species via satellite.

Marais et al. (2018) obtained mean mid-tropospheric NO_2 mixing ratios for various seasons using variations in partial OMI NO_2 columns as a function of optical cloud height (i.e., cloud-slicing). Comparing the cloud-slice based mixing ratios to model output from GEOS-Chem, they inferred a PE of 280 mol per flash for both the tropics and midlatitudes (see Figure 18). In a follow-up study, Horner et al. (2024) obtained higher spatial resolution mean NO_2 mixing ratios for several vertical layers using TROPOMI retrievals for the years 2018–2022. They found that GEOS-Chem mixing ratios exceeded cloud-sliced mixing ratios by 20–50 pptv in the 600–450 hPa layer over the northern midlatitudes during summer north of 35° and attributed the overestimation of NO_2 mixing ratios to the use of a high PE of 500 mol per flash for LNO_x PE in that region.

Kelley et al. (2020) parameterized flash rates in the GISS-E2-1 model in terms of cloud top height while tuning to ensure that total flash rates at both land and marine locations match the OTD/LIS climatology. Constraining the PE using methane lifetime, they found that a PE of 290 mol N per flash (6.4 Tg N a^{-1}) provided the most realistic methane lifetime (9.7 years).

3.6. Discussion of LNO_x PE Estimates

Schumann and Huntrieser (2007) found the most likely annual global LNO_x production to be $5 \pm 3 \text{ Tg N yr}^{-1}$. The mean value was simply an arithmetic mean of all the studies that they reviewed. The range was determined by two

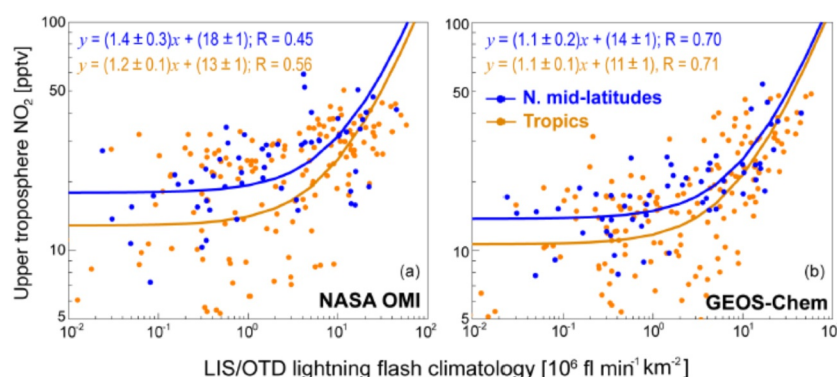


Figure 18. Log-log relationship between upper troposphere NO₂ and lightning flash frequencies, and linear regression fits between the two. Individual points are coincident seasonal mean UT NO₂ from OMI (a) and GEOS-Chem (b) versus seasonal mean LIS/OTD lightning flash climatology for coincident 20° × 32° grid squares in the northern midlatitudes (>30°N; blue) and tropics (<30°N; orange). Northern midlatitude points exclude December–February, which show poor correlation with lightning flashes. The regression lines plot as curves on the log-log scale, highlighting the NO₂ background at low lightning flash rates and the correlation of NO₂ with lightning at high flash rates. From Marais et al. (2018).

methods: using the range of best fit LNO_x production in global models and a range determined from perceived differences in LNO_x production in midlatitude and tropical lightning (from a combination of lightning peak current observations and GEOS-Chem simulations). In our updated analysis we again use the arithmetic mean of all study results considered to obtain the most likely mean annual production, but use a different approach to determine the range because satellite observations have indicated no difference between LNO_x production by a tropical flash and a midlatitude flash (Allen et al., 2019; Bucsela et al., 2019; Marais et al., 2018) and we do not believe the definition of the range should be limited to global model results.

A wide range of LNO_x PE estimates have been published in the 18 years since the review of Schumann and Huntrieser (2007) appeared. The full range extends from 24 to 691 mol fl⁻¹ (a factor of nearly 30). All PE estimates discussed in Section 3.1 through Section 3.5 are illustrated in Figure 19. A large expansion in the number of PE estimates based on NO₂ satellite observations has occurred, allowing a multitude of storms to be sampled. Global model-based estimates have also become more common, and they are often constrained by satellite trace gas observations. We think that satellite-based estimates covering large-scale regions and global model estimates should be considered as the strongest indicators of PE since they account for the effect of a multitude of storms,

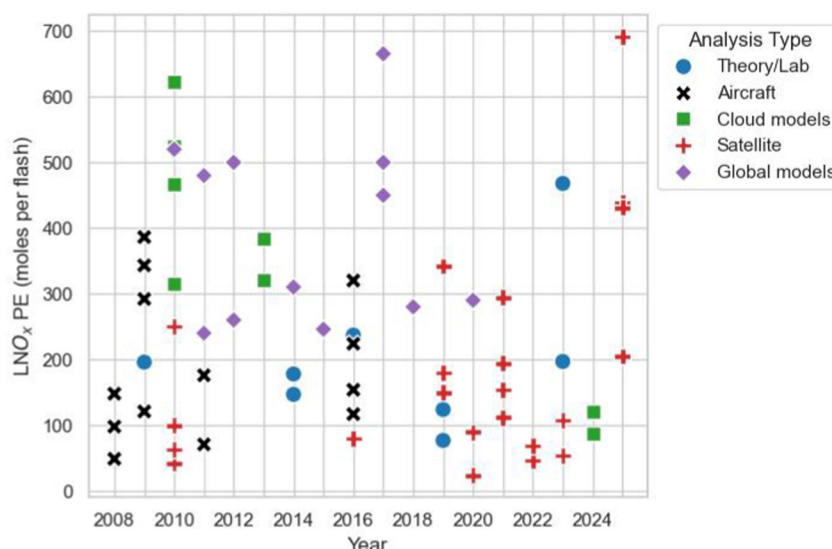


Figure 19. Display of all LNO_x PE estimates presented in Section 3.1 through Section 3.5. PE values are plotted as a function of publication year. Note: LNO_x PE ~250 mol fl⁻¹ is equivalent to global LNO_x production of ~5 Tg N yr⁻¹.

whereas the aircraft data and the cloud-resolved models only consider relatively few cases. Therefore, we use such satellite and global model estimates to bracket the most likely range of actual LNO_x PE.

The satellite observations have yielded LNO_x PE estimates ranging from 24 to 691 mol fl^{-1} , with a mean value of $187 \pm 161 \text{ mol fl}^{-1}$, which is equivalent to $3.7 \pm 3.2 \text{ Tg N yr}^{-1}$. Considering only satellite-based studies covering large-scale regions that use OMI or TROPOMI data (Allen, Pickering, Bucsela, et al., 2021; Allen et al., 2019; Bucsela et al., 2019; Zhang et al., 2020), the mean PE is $145 \pm 38 \text{ mol fl}^{-1}$ or $2.9 \pm 0.8 \text{ Tg N yr}^{-1}$. Although coming from individual storms, mean PE from aircraft observations ($192 \pm 106 \text{ mol fl}^{-1}$, with a range from 49 to 386 mol fl^{-1}) is not much different from that of the satellite observations, yielding an estimate of $3.8 \pm 2.1 \text{ Tg N yr}^{-1}$. The most comprehensive analysis of PE based on the combination of theoretical considerations, laboratory results, and flash observations is that of Koshak (2014). Assuming an IC/CG flash ratio of 3, the mean PE based on the Koshak et al. results for IC and CG flashes yields an overall mean of 178 mol fl^{-1} , equivalent to 3.6 Tg N yr^{-1} . Therefore, the satellite, aircraft, and theoretical/laboratory methods all yield relatively similar mean LNO_x PE values.

Model-based LNO_x PE estimates tend to be considerably larger than those from other approaches. Global and regional models produced PE values ranging from 240 to 665 mol fl^{-1} with a mean of $374 \pm 117 \text{ mol fl}^{-1}$, equivalent to $7.5 \pm 2.3 \text{ Tg N yr}^{-1}$. PE estimates from cloud-resolving models for individual storms range from 88 to 623 mol fl^{-1} , with a mean of $355 \pm 174 \text{ mol fl}^{-1}$, equivalent to $7.1 \pm 3.5 \text{ Tg N yr}^{-1}$. It remains to be determined why both types of models yield PE values on the higher end of the spectrum compared with observation-based estimates, even as most of the model PE estimates are constrained by either satellite or aircraft observations. Possible explanations include error in the specification of background and/or air mass factor in the satellite analyses, unrepresentativeness in aircraft observations, and error in placement of LNO_x emissions or missing chemical reactions in the models.

Based on the above analysis, averaging the mean LNO_x PE values for each type of PE study we obtain an overall mean of 257 mol fl^{-1} , equivalent to 5.1 Tg N yr^{-1} . Considering only the satellite analyses covering large domains, the overall mean becomes 249 mol fl^{-1} , equivalent to $\sim 5 \text{ Tg N yr}^{-1}$. Therefore, the most likely mean annual global LNO_x production is approximately $5 \pm 2 \text{ Tg N yr}^{-1}$ ($3\text{--}7 \text{ Tg N yr}^{-1}$ coming from the observational and modeling studies, respectively). A majority (55%) of all PE results from the aircraft, satellite, and global/regional modeling studies considered in this analysis fall within this range. This result represents a small narrowing of the Schumann and Huntrieser (2007) range of $5 \pm 3 \text{ Tg N yr}^{-1}$ ($2\text{--}8 \text{ Tg N yr}^{-1}$). Only 35% of the studies considered by Schumann and Huntrieser (2007) are contained within the 3 to 7 Tg N yr^{-1} range, thereby providing strong support for the narrowing of the range.

4. Production of Other Trace Gases (O_3 , N_2O , OH, and HO_2) by Electrical Activity in Storms

Several trace gases, including the oxidizing agents ozone (O_3), hydroxyl radical (OH), and hydroperoxyl radical (HO_2), are *directly produced* as part of lightning discharges. Lightning-produced oxidants have a strong impact on the chemical composition of the upper troposphere, due to their high reactivity, and are important drivers of climate change (Chameides, 1986; Lelieveld et al., 2016; Mao et al., 2021). The *indirect production* of trace gases through LNO_x is discussed in Section 5, see also for example, Murray et al. (2013). The direct production of O_3 and nitrous oxide (N_2O) is discussed in Section 4.1. N_2O is a strong greenhouse gas like O_3 but the least reactive oxide of nitrogen. The direct production of OH and HO_2 is discussed in Section 4.2. The direct production of CO, HNO_2 , and H_2O_2 by electrical discharges is not discussed with the reader being referred to Table 11 of Schumann and Huntrieser (2007).

4.1. Direct Production of O_3 and N_2O by Corona and Subvisible Discharges

Contrary to NO, which is produced by hot or flash discharges, O_3 is primarily produced by cold discharges such as corona but is also produced minimally by large or small flash (spark) discharges (Cooray & Rahman, 2005; Franzblau, 1991; Hill et al., 1988; Levine et al., 1981; Peyroux & Lapeyre, 1982; Rehbein & Cooray, 2001; Simek & Clupek, 2002; Y. Wang et al., 1998). Corona discharges occur at low energies and temperatures before the electric field within the conductive region is broken down by intense flash discharges. The discharges are concentrated in a sheath around streamers that connect opposite charged areas (Coppens et al., 1998; Griffing, 1977). Corona discharges also form on cloud water droplets or ice particles within an electrified cloud

(D. Petersen et al., 2015; Shlanta & Moore, 1972; Sisterson & Liaw, 1990) and have even been identified on tree leaves. For example, Figure 1 in Jenkins et al. (2022) shows a photo of an elongated purple-blue point corona discharge. The production of O_3 by coronas is generated by dissociation of O_2 molecules through the attack of electrons (Eliasson et al., 1987) with the production efficiency varying by polarity with negative discharges producing more O_3 than positive ones (Hill et al., 1988).

The first ground-based evidence of sharp, transient peaks in O_3 production possibly related to corona discharges within thunderclouds was observed by Orville (1967) using spectroscopic techniques. Later, Shlanta and Moore (1972) used grounded sharp metal rods combined with tethered balloon-borne measurements and found transient O_3 spikes of 80–200 ppbv O_3 at 3–5 km altitude beneath thunderclouds. This discovery initiated several laboratory experiments and theoretical estimates of O_3 production rate, see Table 11 in Schumann and Huntrieser (2007) for an overview. The measurements are difficult because low-temperature corona discharges near thunderstorms can also be triggered by the O_3 instruments themselves, on the aircraft fuselage, and at the air sampling inlets, producing mostly multi-second O_3 peaks of ~ 10 –100s of ppbv. These peaks are likely artifacts and not related to natural lightning processes (Ridley et al., 2006; Schumann & Huntrieser, 2007; Sisterson & Liaw, 1990; Zahn et al., 2002).

During the airborne DC3 field experiment, performed in Colorado and Oklahoma in 2012, O_3 amounts were enhanced by ~ 1 –10s of ppbv in anvils due to the production of O_3 by subvisible discharges (Brune et al., 2021) reproduced by subsequent laboratory experiments (Jenkins et al., 2021). In visible laboratory sparks, Jenkins et al. (2021) observed O_3 peaks of up to ~ 300 ppbv. See Section 4.2 for more discussion of subvisible discharges.

Several sources contribute to O_3 amounts observed in clouds. They include production by electrical discharges and rapid vertical transport of O_3 -rich air from the boundary layer triggered by strong deep convection (Lelieveld & Crutzen, 1994; Poulida et al., 1996; Winterrath et al., 1999). More recently, clear evidence of wrapping of O_3 -rich stratospheric air around overshooting thunderstorms has been found by ozone LiDAR and in situ measurements on aircraft (Huntrieser et al., 2016a, 2016b; Pan et al., 2014). Also, electrically generated O_3 is not inert and may be converted to NO_2 in the presence of NO . Therefore, it is difficult to assign high amounts of O_3 observed in active thunderclouds entirely to the production by electrical discharges. All these mentioned processes are present at the same time in active thunderclouds and therefore difficult to separate.

Quantification of direct O_3 production by corona discharges is needed to explain the broader O_3 enhancements measured in thunderclouds (Shlanta & Moore, 1972; Winterrath et al., 1999; Zahn et al., 2002). Several follow-up field experiments, using balloons equipped with ozonesondes and in situ instruments on aircraft (Bozem et al., 2014; Kotsakis et al., 2017; Minschwaner et al., 2008) have confirmed the importance of corona discharges as a source of O_3 . In these studies, the contributions of corona discharges to O_3 enhancements in thunderclouds ranged from 12 to 38 ppbv (Bozem et al., 2014; Minschwaner et al., 2008; Winterrath et al., 1999) to as much as 70–100 ppbv (Kotsakis et al., 2017). Table 8 contains a post-Schumann & Huntrieser (2007) overview of O_3 production rate estimates by electrical discharges.

Until recently, the contributions of direct lightning emissions to the global O_3 budget was believed to be minor. Schumann and Huntrieser (2007) report a 2% contribution, while Minschwaner et al. (2008) attribute 40–180 Tg O_3 yr^{−1} to global thunderstorm corona discharges. Compared to O_3 production by lightning-produced NO_x , the given value is a factor of 5 lower and amounts to only $\sim 3\%$ of the total chemical production rate of tropospheric O_3 globally (given as 3,000–4,600 Tg O_3 yr^{−1}, according to Jacob, 1999). More recently, estimates based on O_3 production by corona discharges indicate a factor of 10 higher contribution in the range of 140–1,400 Tg O_3 yr^{−1} globally (Mao et al., 2021), which is of the same order as the stratosphere-troposphere flux of O_3 , which is believed to be 400–1,100 Tg O_3 yr^{−1} (Archibald et al., 2020; Griffiths et al., 2021; Jacob, 1999). As a result, the production of O_3 by corona discharges is now recommended to be implemented into chemical transport models (Kotsakis et al., 2017).

Corona discharges also produce N_2O (Bhetanabhotla et al., 1985; Brandvold et al., 1989; Donohoe et al., 1977; Hill et al., 1984, 1988). See Table 8 for an overview of N_2O production rates from the literature. When discharges occur, N_2O can be generated by the excited state of N_2 (Eliasson et al., 1987), by a metastable variation of the N_2 molecule (Zipf, 1980), and possibly by other reactions discussed in Smith et al. (2021). Airborne in situ measurements conducted by Levine and Shaw (1983) delivered the first evidence of N_2O production by corona discharges in the field, where N_2O spikes in the range of ~ 400 –700 ppbv were observed. Later Brandvold

Table 8
Production Rates for O_3 , N_2O , OH , and HO_2 Produced Directly by Electrical Discharges in Thunderstorms

Species	Value reported in reference	Global emission rate, $Tg\ a^{-1b}$	Reference	Method
O_3				
		0.0005–33	range given in review by Schumann and Huntrieser (2007)	Theory
		6–300	range given in review by Schumann and Huntrieser (2007)	Laboratory
	$4.02 \times 10^{37}\ mol\ a^{-1}$ corresponds to: 2.90×10^{28} molecules per flash (incl. thunderstorm-induced stratospheric intrusions)	3,200 (incl. thunderstorm-induced stratospheric intrusions)	Jadhav et al. (1996)	Field (ground-based spectrometer)
	mean: 1×10^{27} range: $(0.35\text{--}1.6) \times 10^{27}$ molecules per flash ^a	mean: 110 range: 40–180 (by corona discharges)	Minschwaner et al. (2008)	Field (balloon-borne ozonesonde)
	mean: 5.12×10^{28} range: $(0.10\text{--}9.82) \times 10^{28}$ molecules per flash ^a	mean: 5,670 range: 110–10,870 (by corona discharges)	Bozem et al. (2014)	Field (aircraft-borne chemiluminescence detector)
	mean: 2.48×10^{28} range: $(0.87\text{--}3.89) \times 10^{28}$ molecules per flash ^a	mean: 2,740 range: 960–4,300 (by corona discharges)	Kotsakis et al. (2017)	Field (balloon-borne ozonesonde)
	mean: — range: $(3\text{--}8) \times 10^{15}$ molecules in total per discharge (visible and subvisible)	mean: N/A range: 140–1,400 (by corona discharges)	Mao et al. (2021)	Calculation based on values given in Minschwaner et al. (2008), Bozem et al. (2014), and Kotsakis et al. (2017)
N_2O			Jenkins et al. (2021)	Laboratory (max. O_3 production observed near transition from subvisible to visible discharges)
		$(100\text{--}500) \times 10^{-6}$	range given in review by Schumann and Huntrieser (2007)	Theory
		$(1.5\text{--}130) \times 10^{-6}$	range given in review by Schumann and Huntrieser (2007)	Laboratory
OH			Hill and Rinker (1981) & Bhetanabhotla et al. (1985)	Theory
	$(3.4\text{--}5) \times 10^{-8}\ mol\ cm^{-3}$ (per spark discharge)		Jenkins et al. (2021)	Laboratory
	$(2.9 \pm 0.7) \times 10^{-14}\ mol\ cm^{-3}$ (per spark discharge) corresponds to: $(1.1 \pm 0.1) \times 10^8\text{--}(1.34 \pm 0.8) \times 10^9\ mol\ cm^{-3}\ s$ (per spark discharge)		Brune et al. (2021)	Theory (field observations of enhanced OH peaks were modeled with photochemical box model)
	$(1.5 \pm 1) \times 10^9\ mol\ cm^{-3}\ s$ (per subvisible discharge)		Mao et al. (2021)	Theory
	$1.2\text{--}11.8 \times 10^9\ mol\ cm^{-3}\ s$ in total per discharge (visible and subvisible)			
	$(3.1\text{--}27) \times 10^{25}\ mol\ s^{-1}$ $\sim 10^8$ molecules cm^{-3} (based on Brune et al., 2021) (subvisible, corona and flash discharges)	3–30 Tmol $OH\ a^{-1}$ (subvisible, corona and flash discharges)		
HO_2			Hill and Rinker (1981), Bhetanabhotla et al. (1985)	Theory
	$9.0 \times 10^{-11}\ mol\ cm^{-3}$ (per spark discharge)		Jenkins et al. (2021)	Laboratory
	$(6 \pm 2) \times 10^{-14}\ mol\ cm^{-3}$ (per spark discharge)			

^aDirect O_3 production by lightning-associated corona discharges. ^bValue stated in the original paper or by assuming a global flash rate of $44\ s^{-1}$ (Christian et al., 2003), or occurrence of 1,800 electrically active convective cells globally every second (Schumann & Huntrieser, 2007), and the molar mass of the respective species.

et al. (1996) used grounded metal tubes with sharp tips to capture both the production of O_3 and N_2O by corona discharges during thunderstorm passages. The contribution of direct lightning emissions to the global N_2O budget is believed to be negligible (0.002%) (Schumann & Huntrieser, 2007).

In summary, corona discharges are the primary mechanism for direct O_3 production in thunderstorms. Recent observations suggest the magnitude of O_3 production by this mechanism may be sufficient to warrant inclusion of this source of O_3 in atmospheric models. N_2O is also produced in coronas, but in very small quantities.

4.2. Direct Production of OH and HO_2 by Different Types of Electrical Discharges

As mentioned earlier, lightning-produced NO *indirectly* initiates the production of OH and HO_2 , the primary oxidants in the atmosphere. These important cleansers of the atmosphere are also *produced directly* by different types of electrical discharges, as will be described next. By the attack of electrons, H_2O molecules dissociate to form OH and in a second step HO_2 (Ershov and Borysow, 1995). OH and HO_2 are collectively known as HO_x , with HO_x formed from electrical phenomena referred to as LHO_x .

Jenkins et al. (2022) studied the process behind LHO_x production and found that ultraviolet radiation (UV), emitted from corona discharges has the potential to break up H_2O and O_3 to form huge amounts of OH through photolysis. Recently it was suggested that UV also initiates LHO_x production in sparks and flash discharges (Jenkins & Brune, 2025) with most OH formed in the cooler corona sheath surrounding the hot lightning channel (Rakov & Uman, 2006).

From several past model and laboratory studies it has been demonstrated that high amounts of LHO_x can be generated by weak electrical coronal discharges and subsequent streamers (Bhetanabhotla et al., 1985; Hill & Rinker, 1981; Ono & Oda, 2002; Ripoll et al., 2014). See Table 8 for HO_x production rates associated with these processes. However, these studies concluded that the lifetime of LHO_x is too short ($\sim$ milliseconds) to be relevant for the global OH budget. Interestingly, the global OH production by corona on trees below electrified clouds, 1×10^{25} – 6×10^{26} mol s^{-1} according to Jenkins et al. (2022) is in the same range as given by Mao et al. (2021) for production by discharges (see Table 8). However, the total production by these processes is only a tiny fraction of the global OH production of $\sim 4 \times 10^{30}$ mol s^{-1} (Ferracci et al., 2018).

The first hints that HO_x concentrations can be surprisingly high in thunderstorm outflow were reported from the airborne GABRIEL field experiment, although the high values were related to enhanced photochemical activity by Bozem et al. (2014). The first evidence of extreme HO_x production in active thunderstorms, attributed to both visible and subvisible lightning, was observed by Brune et al. (2021) during DC3 (see Figure 20). LHO_x varied from ~ 100 to ~ 1000 s of pptv (max. $\sim 2,000$ pptv), values which exceeded previously reported atmospheric values by 1–2 orders of magnitude (Brune et al., 2018; Stone et al., 2012). Subsequent laboratory experiments by Jenkins et al. (2021) (see also Brune et al., 2021), clearly demonstrated that both visible and subvisible discharges play an important role for the formation of HO_x but not the formation of NO or O_3 . The importance of invisible discharges was somewhat surprising as these discharges are much weaker and less charged compared to lightning sparks and corona discharges. These laboratory findings also clarified why, during DC3, high HO_x peaks appeared in the electrified anvil region where no visible lightning was detected. In a follow-up study by Brune et al. (2022), it was demonstrated that extensive amounts of OH and HO_2 (HO_x exceeding 3,000 pptv, highest atmospheric value ever observed) can also be produced by visible corona discharges on grounded metal objects below thunderstorms. After analysis, it was concluded that the OH and HO_2 peaks observed during DC3 were most likely of atmospheric origin and not an artifact of corona discharges on the aircraft sampling inlet.

First quantifications of the role LHO_x plays in the global OH budget based on the results from Brune et al. (2021) indicate contributions in the range of 2%–16%. This means that thunderstorms with all their types of discharges: flash discharge, sparks, streamers, corona and subvisible discharges, play a non-negligible role for the global HO_x budget. Mao et al. (2021) was the first to include this source in a model. Simulations by Mao et al. indicate a highly uncertain contribution from lightning-produced oxidants in the range of 0.3%–10% to global mass weighted OH, in general agreement with Brune et al. (2021) findings.

Recently the discovery of subvisible discharges has been of large interest, since it is hypothesized that they play an essential role for the initiation of lightning (Kostinskiy et al., 2020; D. Petersen et al., 2015; Rison et al., 2016). However, only a minority of all present weak discharges form a lightning flash. Because the weaker discharges

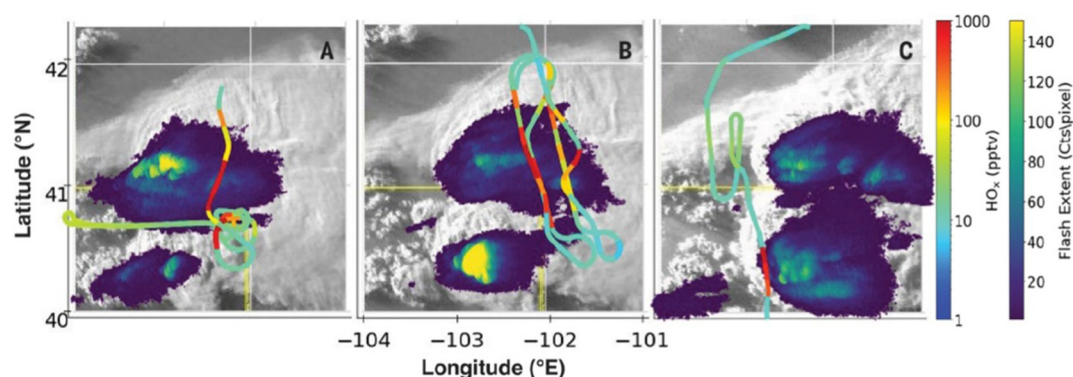


Figure 20. HO_x abundances (indicated by colors and logarithmic scale) along the DC-8 flight track overlaid on LMA flash extent and satellite anvil images during the flight on 22–23 June 2012 for three 1-hr periods: (a) 23:30–00:30 UTC; (b) 00:30–01:30 UTC; and (c) 01:30–02:30 UTC. The flash extent locations are shown for the entire hour as the storms moved from west to east. The storm cores are regions of the most intense flash extent. Cts/pixel, counts per pixel (each pixel is about 1 km^2). From Brune et al. (2021).

are much more frequent and cover a larger volume in the thunderstorm, it is assumed that they have the same relevance for LHO_x as normal flash discharges (Jenkins et al., 2021).

Finally, the recent discovery of so-called Blue Luminous Events (BLUES), a common transient corona discharge in the upper part of thunderclouds ($\sim 8.5\text{--}15 \text{ km}$ altitude), suggests that such discharges may produce considerable but highly uncertain amounts of N_2O , O_3 , OH , and HO_2 (Chanrion et al., 2017; Soler et al., 2021). BLUES are related to positive and negative narrow bipolar events (NBEs), have a short duration of a few to hundreds of milliseconds, and are typically observed in a narrow (not visible to the human eye) blue band at 337 nm (F. Liu et al., 2018; Rison et al., 2016; Soler et al., 2020). The global rate of BLUES has been determined to be $\sim 3.5 \text{ per s}$ (Soler et al., 2022), which is 12 times less frequent than normal lightning (Blakeslee et al., 2020). Such corona discharges have the capability to excite common atmospheric molecules such as N_2 , O_2 , and H_2O by non-thermal equilibrium electron-impact collision to form N_2O , O_3 , OH , and HO_2 (Gordillo-Vázquez et al., 2018). Perturbations to these gases, especially in the climate sensitive upper tropospheric and lower stratospheric region, have a strong impact on the radiation balance. Therefore, the influence of BLUES on atmospheric composition needs more attention (Gordillo-Vázquez & Pérez-Invernón, 2021) especially in regions where a high density of sub-visible discharges and BLUES have been observed, as in e.g. the Tornado Alley in North America (Brune et al., 2021; Soler et al., 2022). Recently, it has been proposed by Soler et al. (2022, 2024), that the presence of such BLUES might explain the extreme production of LHO_x reported by Brune et al. (2021) and Jenkins et al. (2021), which they related to so-called “subvisible” discharges, and also the short-term extreme production of O_3 in thunderclouds reported on in the first paragraphs of this section.

LHO_x is primarily produced in corona and subvisible discharges. The quantities produced in both laboratory experiments and in observed storms may be large enough to warrant inclusion in atmospheric chemical models.

5. Photochemical Production of O_3 Downwind of Storms

Ozone concentrations downwind of thunderstorms vary widely due to the rapid vertical mixing of boundary layer (BL) and upper tropospheric (UT)/lower stratospheric (LS) air masses (Thompson et al., 1997). Huntrieser et al. (2016a, 2016b) investigated the mixing of these air masses due to processes associated with deep convection over the US during the DC3 experiment. A combination of trace gas data, radar, lightning, satellite retrievals and model results were used to analyze various transport pathways associated with thunderstorms and the impacts on the upper troposphere and lower stratosphere (UT/LS) O_3 distributions. Overshooting cloud tops transported tropospheric air from rural locations that had low amounts of O_3 and substantial LNO_x into the LS. As part of the same system, high O_3 air from the LS was transported downward into the anvil and the region surrounding the outflow. With minimal photochemical production of O_3 in the freshly lofted air, the O_3 distributions observed in these storms were determined by dynamical processes in agreement with Pan et al. (2014).

Over time, photochemical production within the thunderstorm outflow produces additional O_3 (Apel et al., 2015; Martini et al., 2011; Pickering et al., 1990), with the magnitude of the enhancement varying with BL air composition and LNO_x PE (DeCaria et al., 2005; Ott et al., 2007; Park et al., 2004; Pickering et al., 1990, 1992). Over and near-land, the BL air often has high concentrations of O_3 precursors or O_3 itself from sources that include anthropogenic emissions and biomass burning, that when convectively lofted along with LNO_x lead to substantial photochemically produced O_3 in the UT downwind of storms. The UT/LS air is often ozone-rich (e.g., Crutzen, 1995; Huntrieser et al., 2016a, 2016b) but can be reduced in O_3 content by lofting of low O_3 BL air from rural areas. Downwind UT photochemical O_3 production leads to restoration of O_3 in this layer (Pickering et al., 2024).

5.1. Photochemical O_3 -Production by LNO_x and Its Impact on Upper Tropospheric O_3

In this section we first examine the large-scale impact of LNO_x on UT O_3 . We then review the results of six studies that estimate ozone production downwind of specific deep convective events (see Table 9).

The magnitude of the large-scale enhancements of UT- O_3 production due to LNO_x has been demonstrated through analysis of ozonesonde data, airborne data and through comparisons of global model output with satellite data. For example, box model calculations by O. R. Cooper et al. (2007) identified LNO_x as the cause of a 25–30 ppbv enhancement in 10-km O_3 observed by ozonesondes over the southeastern US over a 10-day summertime period. Martin et al. (2007) obtained an increment of 10–15 Dobson Units of tropospheric O_3 downwind of convection with impacts of LNO_x and biomass burning through analysis of their GEOS-Chem model output constrained with SCIAMACHY NO_2 and OMI/MLS O_3 satellite data.

Huntrieser et al. (2011) examined O_3 production by airborne in situ measurements in the outflow of mesoscale convective systems (MCS) observed during the AMMA campaign. In the presence of aged biomass burning plumes, they found unusually high O_3 production rates of up to 30 ppbv within 2–3 hr of outflow transport. Morris et al. (2010) noted an increase in O_3 of 1.6–4.6 ppbv hr^{-1} in the 2.5–5.1 km layer from an ozonesonde that had oscillated up and down five times in the edge of a dissipating convective cell originating over the Gulf of Panama during the TC⁴ mission.

Apel et al. (2012) noted from a box model calculation a net O_3 increase of ~ 10 ppbv between 9 and 12 km altitude over a few hours following observed convection with lightning over Canada in the Arctic Research of the Composition of the Troposphere from Aircraft and Satellite (ARCTAS) experiment. They attributed the high O_3 production rate to the combination of LNO_x and convective mixing that brought unusually large concentrations of isoprene from the boundary layer to the mid- and upper troposphere. Unfortunately, here no downwind flights were able to verify the O_3 production shown by the model. Later, Apel et al. (2015) examined the O_3 production downwind of two Colorado storms during the DC3 experiment on 22 June 2012, a time period when the High Park wildfire was active in the area. Observations by aircraft showed that the northern storm had larger VOC reactivity than the southern storm due to its wildfire smoke, while the southern storm had higher flash rates and LNO_x production. Their model results indicated that 11 (14) ppbv of O_3 were produced downwind of the northern (southern) storm (Figure 21). The larger O_3 production of the latter storm was caused by LNO_x as indicated by modeling, since only 2 ppbv O_3 was produced when the LNO_x source was not included.

Bozem et al. (2014) presented results from the GABRIEL airborne campaign based in Suriname in October 2005 (dry season). Convective influence was investigated in outflow between 9.5 and 11 km altitude from a storm on October 5th. NO was enhanced in the outflow by a factor of 2–3, likely due to lightning observed by the LIS. Photochemical O_3 production in the outflow was calculated to be 1.7–4.8 ppbv/day. Pickering et al. (2024) examined the O_3 production downwind of a 29–30 May 2012 storm that also occurred during the DC3 experiment. Aircraft sampling showed that O_3 amounts in the outflow between 10.7 and 12.2 km altitude increased from 86 ppbv the day of the storm to 105 ppbv the following day over the Southern Appalachians, an increase of 19 ppbv. Subsequent model simulations (Figure 22) indicated that 10–11 ppbv of the 24-hr increase was due to photochemical O_3 production impacted by LNO_x . This magnitude of O_3 production in the 24 hr following a storm was similar to that computed for a Colorado storm in the STERAO-A experiment by DeCaria et al. (2005).

Nussbaumer et al. (2023) used several quantities including the fraction of methyl peroxy radicals forming formaldehyde $\alpha(CH_3O_2)$ and the ratio of HCHO to NO_2 as metrics to determine the dominant chemical regime in

Table 9
Estimates of Photochemical Production of Ozone Due To LNO_x Downwind of Deep Convection

Reference	Experiment	Region	Method	UT P(O ₃)
Pickering et al. (1996) ^a	TRACE-A	Brazil	Aircraft obs and 1-D chemistry model	8–10 ppbv in 24 hr
DeCaria et al. (2005) ^a	STERAO-A	Colorado	Aircraft obs and cloud-resolved chemistry model	10 ppbv in 24 hr
Ott et al. (2007) ^a	EULINOX	Germany	Aircraft obs and cloud-resolved chemistry model	−2 to +5 ppbv in 24 hr (polluted conditions)
Cooper et al. (2007)	INTEX-A	Houston, TX	Ozonesondes and box model	25–30 ppbv over 10 days
Martin et al. (2007)	–	Tropical Atlantic and Africa	GEOS-Chem model constrained by satellite data	10–15 DU
Morris et al. (2010)	TC ⁴	Gulf of Panama	Ozonesonde	1.6–4.6 ppbv per hour at edge of dissipating cell
Huntrieser et al. (2011)	AMMA	West Africa	Aircraft observations	Up to 30 ppbv in 2–3 hr
Apel et al. (2012)	ARCTAS	Canada	Aircraft obs and box model	10 ppbv over a few hours
Bozem et al. (2014)	GABRIEL	Suriname	Aircraft obs and photochemical production and loss calculations	1.7–4.8 ppbv per day
Apel et al. (2015)	DC3	Colorado	Aircraft obs and box model	11–14 ppbv over 2 days
Pickering et al. (2024)	DC3	Oklahoma	Aircraft obs and WRF-Chem model	10–11 ppbv in 24 hr

^aThis reference is included in Schumann and Huntrieser (2007) and is provided here for comparison with more recent estimates.

the upper troposphere, a region where LNO_x is usually the predominant source of NO_x. They found that lightning is the major determinant of whether ozone chemistry is NO_x- or VOC-sensitive.

Gressent et al. (2014) examined the impact of lightning on large-scale (>300 km) upper tropospheric plumes of NO_y using MOZAIC measurements over the North Atlantic region during 2001–2005. After removing air masses affected by aircraft emissions or stratosphere-troposphere exchange, backward trajectories to the locations of deep convection and lightning were used to extract lightning-influenced NO_y plumes. Analyzing the large-scale lightning-influenced plumes, they found that the majority originated from extratropical cyclones over North America that were transported over the Atlantic via warm conveyor belts. The transport resulted in a negative west-to-east gradient in NO_y and a positive west-to-east gradient in O₃.

Pérez-Invernón et al. (2025) examined the change in atmospheric composition obtained when the LNO_x PE in EMAC (Jöckel et al., 2016) was allowed to vary (decrease) with flash rate as in Bucselá et al. (2019). The resulting more spatially homogeneous distribution of LNO_x led to subsequent changes in atmospheric composition including more O₃ over the tropical ocean and less over tropical Africa and a few percent decrease in the lifetime of tropospheric OH and concentrations of CO and HNO₃.

Ghosh et al. (2025) added LNO_x production to the regional CHIMERE model and examined the sensitivity of atmospheric composition to the LNO_x source via three Northern Hemisphere simulations at 100 km horizontal resolution. With LNO_x PE set to 697 mol fl^{−1} for CG flashes and 199 mol fl^{−1} for IC flashes, total hemispheric LNO_x production was 2.8 Tg N yr^{−1} when a cloud-top height scheme was used to parameterize flash rates and 3.1 Tg N yr^{−1} when an ice flux-based scheme was used. Compared to a run with no lightning, the greatest O₃ increases (6–10 ppbv) appeared in the 750–500 hPa layer over the tropics. The addition of LNO_x also resulted in a 7%–11% increase in the hemispheric O₃ burden and a 11%–17% reduction in CH₄ lifetime.

Finney, Doherty, Wild, & Abraham (2016) examined the impact of lightning-NO_x on the odd oxygen (O_x) budget using the cloud-height and ice flux schemes within the UK Chemistry and Aerosols Model (UM-UKCA). For a simulation representative of year 2000 conditions, they found that lightning-NO_x was responsible for 20%–25% of the tropospheric O_x burden depending on the scheme, ~19% of its chemical production and loss, and ~11% of its deposition. LNO_x also reduced the lifetime of O₃ by ~6%. Figure 11b of Finney et al. shows the mean column O_x

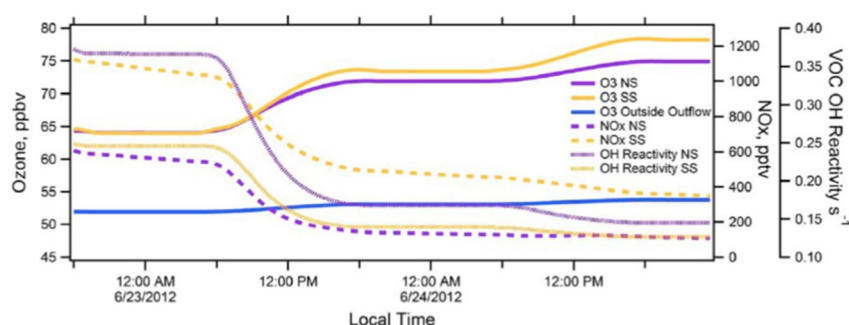


Figure 21. Time series showing mixing ratios of O_3 (ppbv), NO_x (ppbv), and OH reactivity (s^{-1}) for the northern (NS) and southern (SS) Colorado storms. The 54-hr simulations were performed using the NCAR master mechanism initialized with mean mixing ratios of species from the respective storms and outside of the outflow. From Apel et al. (2015).

production per continental flash as a function of flash rate for the cloud height and ice flux schemes. Initial O_3 production per flash was found to be ~ 10 times greater for low flash rates than high flash rates with the ice flux scheme showing the greatest contrast in efficiency between low- and high-flashing cells.

Nalam et al. (2025) simulated global tropospheric O_3 using tagged NO_x and reactive carbon tracers in the CAM4-Chem model run at $1.9 \times 2.5^\circ$ resolution driven by MERRA-2 meteorology. LNO_x emissions were set at $3.47 \text{ Tg N yr}^{-1}$. While lightning was the second largest contributor (22%) to the tropospheric O_3 burden behind anthropogenic sources (37%), it was the most efficient at producing O_3 ($5.93 \text{ mol } O_3 \text{ per mol } NO_x$) as a large fraction of LNO_x production occurred aloft where concentrations of NO_x were lower.

To summarize, downwind of electrically active storms LNO_x leads to photochemical O_3 production in the UT at rates most commonly in the range of 5–10 ppbv per day. Larger rates have been observed over a few hours in the near field of convection.

5.2. Photochemical O_3 -Production by LNO_x and Its Impact on Surface O_3

The effects of lightning- NO_x (LNO_x) on surface O_3 over the United States have been investigated in several studies using the US Environmental Protection Agency Community Multiscale Air Quality (CMAQ) model:

- Kaynak et al. (2008) examined the impact of adding an LNO_x source on daily maximum 8-hr surface O_3 ($8hrO_3$). Assuming a production efficiency (PE) of 500 mol of NO_x per flash, they found that changes in $8hrO_3$ due to LNO_x were < 2 ppbv in 71% of grid cells with a maximum increase of 10 ppbv.

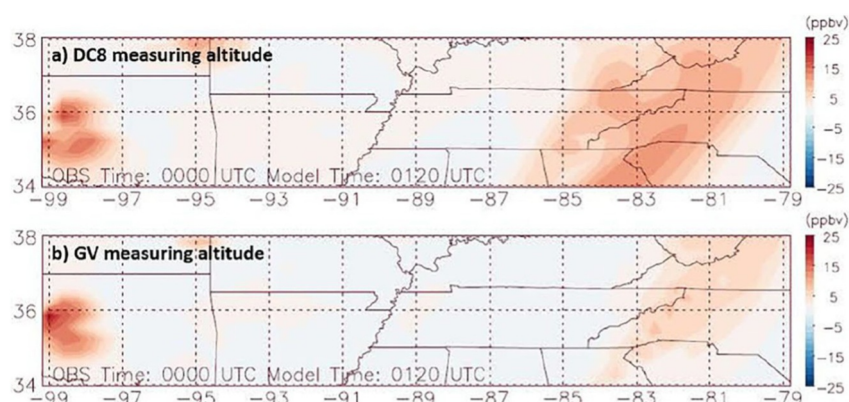


Figure 22. 24-hr simulated net ozone production between 00:00 UTC 30 May (model time: 01:20 UTC) and 00:00 UTC 31 May (model time: 01:20 UTC) in air arriving at map locations at the mean DC-8 altitude of 11 km (a) and mean GV altitude of 11.5 km (b). Note air with maximum net ozone production arrived in a band from southern West Virginia (WV) to northern Georgia (GA). From Pickering et al. (2024).

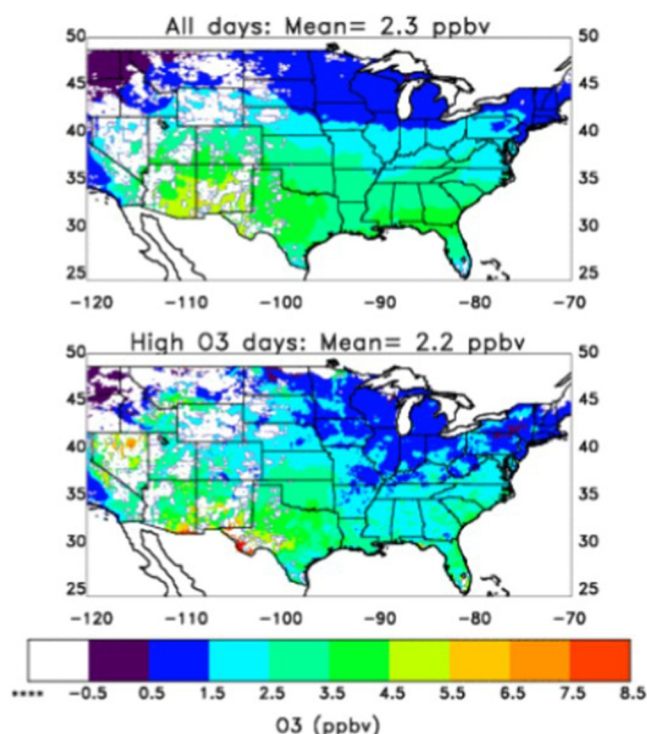


Figure 23. Mean increase in 8-hr maximum ozone associated with lightning-NO production during the 1 June through 31 August time period. Results averaged over all days (top) and over days when 8-hr maximum ozone from simulation with no lightning exceeds 60 ppbv (bottom). From Allen et al. (2012).

- Allen et al. (2012) also implemented a LNO_x scheme in CMAQ. It assumed flash rates were proportional to model convective precipitation but constrained locally such that monthly average flashes matched an NLDN-based estimate of total flashes. For a 500 mol per flash source, tropospheric NO₂ columns in summer 2006 were reproduced within −5% to +13% of OMI observations over CONUS. The contribution of LNO_x to 8hrO₃ was found to be 1.5–4.5 ppbv. See Figure 23.
- Kang et al. (2020) estimated total flash rates using observed hourly NLDN CG flashes and climatological IC/CG ratios (Boccippio et al., 2001). Using a PE of 350 mol fl^{−1}, CMAQ produced increases in 8hrO₃ as large as 17 ppbv over the US mountain west region. The increase in 8hrO₃ over this region due to LNO_x averaged 6 ppbv and resulted in improved model performance.
- P. Cheng et al. (2024) and W.-Y. Cheng et al. (2024) used the Beta methodology described by Wu et al. (2023) to include LNO_x production in CMAQ. In this scheme, optical energy from GLM was used to introduce flash-by-flash variations in LNO_x production. Assuming a mean PE of 250 mol fl^{−1}, they found only small changes in surface (1 ppbv over southeast US) and column O₃ (1%–2% changes) during the June–September 2019 time period.

The WRF-Chem model has also been used to estimate the LNO_x impact on surface O₃. Li et al. (2022) used WRF-Chem to simulate the impact of LNO_x on surface O₃ on the Tibetan Plateau. They parameterized flash rates in terms of Price and Rind (1992) defining the cloud top as the level of neutral buoyancy as in Wong et al. (2013). Vertical distributions of LNO_x were specified using the profiles of Ott et al. (2010). LNO_x PE values ranging from 0 to 330 mol fl^{−1} were tested, and they found that a PE of 200 mol fl^{−1} best fit observations. The contribution of lightning to summertime 8hrO₃ was 9.3 ± 7.1 ppbv over the Tibetan Plateau, representing $17.5 \pm 14.5\%$ of the total O₃ mixing ratios.

Menut et al. (2020) computed the impact of LNO_x on surface O₃ over Europe and northern Africa using the WRF-CHIMERE set of models. Flash rates were predicted from the model cloud top heights (Price & Rind, 1992), and the LNO_x emission profile was constant with altitude. The Price et al. (1997) values of PE for CG flashes of 1,100 and 110 mol fl^{−1} for IC flashes were used. Over most of Europe a <1 ppbv contribution of lightning to surface O₃ was computed, with a few patches of 1–3 ppbv. The maximum contributions of 10 ppbv were found over Africa in the 10–15°N band.

LNO_x production also contributes to deposition of reactive nitrogen to the surface. Shepon et al. (2007) examined the deposition of LNO_x using The Model of Atmospheric Transport and Chemistry-Max Planck Institute for Chemistry (MATCH-MPIC). They found that approximately 50% of LNO_x was deposited over the oceans even though most LNO_x is produced over land. The large oceanic component is a consequence of the long lifetime of LNO_x, its sequestering as PAN or HNO₃, and strong upper tropospheric winds. They note that an improved understanding of this term is essential to estimate how much carbon is sequestered as the result of LNO_x deposition. Allen et al. (2012) and Kang, Foley, et al. (2019) showed that the inclusion of LNO_x in CMAQ generally improved wet deposition and ambient concentrations of O₃ and NO_x.

Regional model simulations with LNO_x show an average impact on surface daily maximum 8-hr surface O₃ of 1–6 ppbv over the US, with maxima reaching 17 ppbv over high altitude terrain of the western US. Over the high terrain of Tibet, LNO_x has been computed to yield $\sim 9 \pm 7$ ppbv O₃ on average. Over Europe, lesser amounts of surface O₃ (<1 ppbv on average with a peak of 3 ppbv) were attributed to LNO_x.

6. Impact of Changes in LNO_x on the Budgets of OH, CH₄, Aerosol, and VSLs and Radiative Forcing

LNO_x influences the quantities of many atmospheric trace gases, as well as aerosols. In this section we discuss the processes by which LNO_x enhances OH, and subsequently the effect of enhanced OH on the lifetime of CH₄ and the quantity of very short-lived substances that are available for transport to the stratosphere. LNO_x also enhances nitrate aerosol concentrations.

LNO_x production leads to the formation of OH, the most important oxidizing species in the atmosphere (Fiore et al., 2024), through the following processes:

1. NO₂ with a lightning source is photolyzed to NO and O. The O atom combines with O₂ to form O₃. NO is free to react with another O atom to regenerate NO₂ allowing the cycle to continue. Throughout much of the troposphere, the primary OH source is the photolysis of O₃ at wavelengths <310 nm, yielding O(¹D), which in the presence of H₂O vapor (Brasseur & Solomon, 2005) produces OH. NO produced by lightning may also react with HO₂ converting HO₂ to OH, thus affecting HO_x partitioning. NO₂ may react with OH in the presence of a third molecule to form HNO₃, which usually leads to the loss of both HO_x and NO_x.
2. A secondary source of importance for global OH is the recycling of peroxy radicals (HO_x) formed by the reaction of OH with reduced and partly oxidized species, including methane, CO and VOCs (Chua et al., 2023). In polluted air, NO_x emissions control the secondary OH production.
3. Finally, the rate at which OH is formed is sensitive to changes in the depth of convection. For example, increases in the depth of convection will lead to more NO transport into the UT where it increases OH through the reaction with HO₂ discussed above (Banerjee et al., 2014).

Increases in lightning are associated with increases in lightning-NO_x production and subsequently increases in OH, HNO₃, PAN, O₃ and decreases in CH₄ (Labrador et al., 2004; Liaskos et al., 2015). Liaskos et al. found that increasing the LNO_x production rate by a factor of 4 (from 123 to 492 mol flash⁻¹) led to tropical upper tropospheric enhancements of greater than 100% in NO_x, OH, HNO₃, and PAN and up to 60% enhancements in upper tropospheric O₃. Banerjee et al. (2014) used time-slice experiments with a CCM that parameterized flash rates using the Price and Rind scheme (1992, 1994) to examine the impact of increases in LNO_x production on atmospheric composition. They found that including methane driven feedbacks reduced increases in the tropospheric O₃ burden due to LNO_x by 5 Tg (from 29 to 24 Tg) for a Representative Concentration Pathway (RCP) 4.5 scenario and by 13 Tg (from 46 to 33 Tg) for a RCP8.5 scenario.

The relative contribution of lightning (~10% currently) to total NO_x emissions is growing slowly as anthropogenic emissions decrease (Murray, 2016). NO_x production from lightning has a disproportionate impact on climate relative to most emissions because it is predominantly released in the upper troposphere. Murray et al. (2021) discussed the factors controlling the lifetime of NO_x and thus atmospheric OH concentrations: The first factor is the injection height of lightning NO_x. Models with LNO_x emitted exclusively in the PBL have less global NO_y and perhaps 20% less OH than models that parameterize its release more realistically (Murray et al., 2014). The second factor is heterogeneous reactions on aerosol particles such as the hydrolysis of N₂O₅ and uptake of HO₂. Disabling these reactions increased global mean OH by 12%. The possible uptake of HNO₃ on dust particles also affects global mean OH. The third factor is the fate of isoprene nitrates. Models that assume isoprene nitrate is lost by deposition have larger global mean OH content than models that allow it to be recycled back to NO_x.

Choi et al. (2009) examined the indirect impact of O₃ production associated with LNO_x on outgoing longwave radiation (OLR) during June–July 2005 using REAM. They found that the impact of LNO_x on OLR and radiative forcing over the eastern U.S. and western Atlantic (up to 0.35 and 0.55 W m⁻², respectively) was up to a factor of three greater than the impact of anthropogenic-NO_x. They noted that the relative impact of LNO_x is likely to increase in the future as anthropogenic NO_x emissions continue to decrease and LNO_x emissions change.

The lifetime of atmospheric CH₄ is approximately 9 years with about 90% of its loss in the troposphere due to reaction with OH (Holmes et al., 2013; Murray et al., 2014; Saunio et al., 2020). Thus, increases in OH due to possible increases in LNO_x would decrease the atmospheric lifetime of CH₄ (Fiore et al., 2006; Wild, 2007) and other variables such as methyl chloroform whose primary loss mechanism is reaction with OH. Ghosh et al. (2025) used the CHIMERE model to show that there is a 40% increase in global OH burden when LNO_x is included in the model, leading to a reduction of CH₄ lifetime of 24%. Finney et al. (2018) simulated year-2000 and

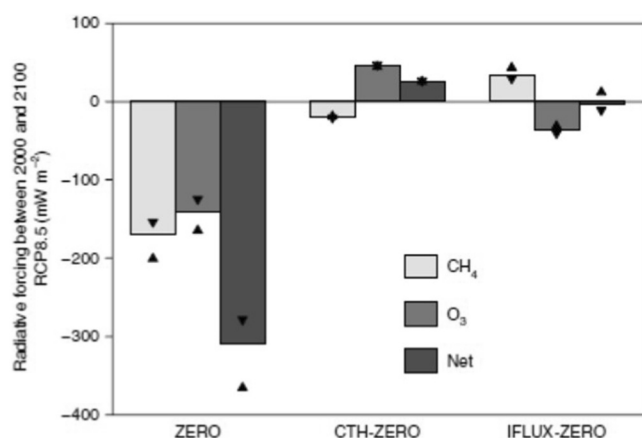


Figure 24. Model estimated changes in year-2000 to year-2100 radiative forcing due to climate-related changes (RCP8.5 scenario) in CH₄, O₃, and “Net” for simulations run without LNO_x (left), with LNO_x and cloud-top height (CTH) based flashes (Price and Rind, 1992) (center), and with LNO_x and ice-flux (IFLUX) based flashes (Finney et al., 2018) (right). From Finney et al. (2018).

year-2100 conditions with the UM-UKCA model with the cloud-height and ice flux lightning schemes. For year-2000 conditions, Finney et al. showed that LNO_x was responsible for a 21% reduction in the CH₄ lifetime and a 6%–8% increase in the O₃ lifetime. By year-2100, the lifetime of O₃ decreased by 16%–18% while the lifetime of CH₄ decreased by 18%–24%.

The increase in H₂O vapor associated with a warming climate will likely be accompanied by a decrease in the tropospheric burden of O₃ (e.g., Unger et al., 2006) with the magnitude of this decrease sensitive to changes in LNO_x. Increases in LNO_x production could negate a major portion of the H₂O vapor induced decreases in tropospheric O₃ burden (Tuomi et al., 1996) at least over time scales of a few years. At longer time scales, increases in LNO_x could lead to a decrease in O₃ burden through its effect on CH₄ (Murray, 2016).

The fraction of very short-lived substances (VSLs; e.g., halogenated organic or inorganic gases) reaching the stratosphere is also sensitive to OH abundance. Increased oxidation of VSLs by OH in the upper troposphere in a future climate with enhanced LNO_x could counteract increased transport of VSLs to the lower stratosphere by more vigorous convection and subsequent advection (Banerjee et al., 2014).

Increases in LNO_x result in increased radiative forcing by O₃ and decreased forcing by CH₄ (Holmes et al., 2013; Iglesias-Suarez et al., 2018; Martini

et al., 2011; Sinha & Tuomi, 1997; Toumi et al., 1996). The O₃ and CH₄ forcing are of similar magnitude but of opposite sign. The CH₄ effect acts on a global scale with a delay of approximately a decade while the O₃ effect is of regional character and occurs on the time scale of weeks (Fuglestad et al., 1999). The radiative efficiency of O₃ with a lightning-source is especially high because LNO_x production is most common at higher altitudes where UV radiances are large, background concentrations of NO_x are small, and the lifetime of O₃ is long (Dahlmann et al., 2011; Lacis et al., 1990).

Khodayari et al. (2018) used CAM5 to examine the sensitivity of increases in O₃ radiative forcing to background lightning-NO_x levels. They found that for a doubling of LNO_x emissions, the northern hemisphere zonal mean O₃ perturbation due to aviation was reduced by a factor of 1.2. Thus, a better understanding of the magnitude of the NO_x sources and sinks from lightning is needed to better quantify the impact of aviation emissions on O₃ production. Finney et al. (2018) used the UM-UKCA with cloud-height and ice-flux based lightning schemes to examine the effects of projected future changes in lightning on radiative forcing by CH₄ and O₃ between 2000 and 2100. The UM-UKCA when run with the cloud-height based flash rate scheme showed a positive lightning-climate feedback as the increase in forcing due to increased O₃ was approximately twice as large as the decrease in forcing due to decreases in CH₄. By contrast, the model when run with the ice flux scheme did not support a lightning-climate feedback as the increase in forcing due to increased CH₄ was balanced by a decrease in forcing due to decreases in O₃ (see Figure 24).

X. Wang et al. (2025) used GEOS-Chem to simulate atmospheric composition over the 2013–2021 period using unconstrained and a Schumann-Resonance (SR)-constrained cloud-height based flash rate schemes. Examining atmospheric composition, the SR-constrained simulation showed a 5.4% per decade decrease in global NO_x concentrations over the period driven primarily by reductions in anthropogenic emissions, which was larger than the 3.8% decrease in the standard simulation, but more in line with observed decreases in anthropogenic emissions. The SR-constrained simulation also did a better job of capturing COVID-related decreases in lightning and subsequent 2% decreases in OH between 2019 and 2020. These decreases are believed to be responsible for approximately 50% of the growth of CH₄ concentrations between 2019 and 2020 (see Figure 25). They noted that capturing interannual variations in flash rates is an essential step when using model output as a guide to interpret the CH₄ budget. Stecher et al. (2025), using a CH₄ emission flux driven chemistry-climate model, found that climate change led to 0.3 Tg N yr^{−1} increase in LNO_x emissions and enhanced OH. They used the multi-model mean sensitivity of CH₄ lifetime to changing LNO_x emissions of −4.8% per Tg N yr^{−1} from Thornhill et al. (2020) to conclude that the 0.3 Tg N yr^{−1} increase would cause a −1.4% change in CH₄ lifetime in their model.

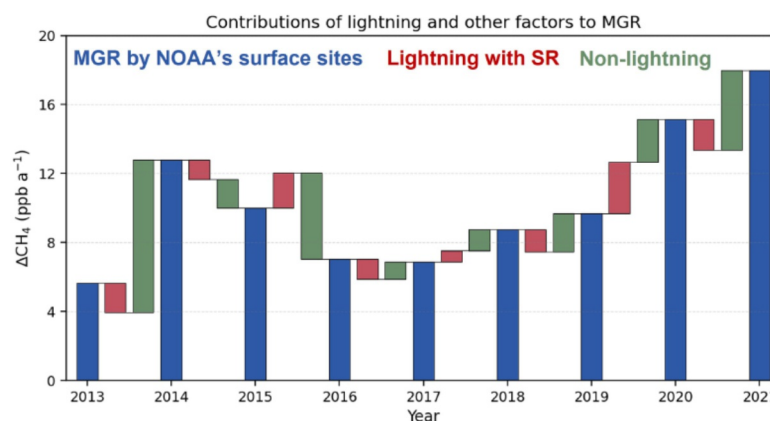


Figure 25. Time series showing the annual atmospheric methane growth rate (MGR) during the 2013–2021 time period (blue) and the contribution of SR-based lightning-driven variations in the methane sink by OH oxidation (red) and non-lightning sources (green) to its interannual variations. The lengths of the red (green) bars show the magnitude of the lightning-induced (non-lightning) variations. It is notable that the contribution of lightning-induced variations in OH to the MGR is usually less than the contribution of non-lightning sources. In 5 of the 8 years, lightning-induced variations oppose non-lightning variations and reduce the change in MGR. However, in 3 years (2014–2015, 2017–2018, and 2019–2020) lightning-induced changes enhance the change in MGR. From X. Wang et al. (2025).

Tost (2017) examined the impact of LNO_x on UT nitrate concentrations using the ECHAM5/MESy chemistry and climate model. With a 6 Tg N yr^{-1} LNO_x source he found that UT nitrate increased by 30% under present day conditions and by 50% in preindustrial conditions when LNO_x was added to the model. Luhar et al. (2024) investigated the LNO_x effect on nitrate aerosol in more detail using the UM-UKCA chemistry and climate model with a 5.2 Tg N yr^{-1} LNO_x source and a modal aerosol nitrate scheme. Addition of LNO_x to the model resulted in global mass burden increases of 4.7% for fine nitrate, 12% for coarse nitrate, 2.8% for ammonium, and 5.8% for sulfate resulting in a 5% decrease in global mean OH, a 4%–5% decrease in global O_3 burden, and a comparable increase in CH_4 lifetime.

Lightning-produced NO reacts with HO_2 to yield OH and NO_2 . NO_2 photolysis leads to photochemical O_3 production as discussed in Section 5. LNO_x also leads to UT increases in HNO_3 and PAN and nitrate aerosol. Enhanced OH resulting from LNO_x production leads to faster oxidation of CH_4 by OH, and thereby decreases the CH_4 lifetime. Sensitivity of CH_4 lifetime to LNO_x has been calculated in global models to be $\sim 4.8\%$ per Tg N/yr change in LNO_x emission. Therefore, increases in LNO_x would lead to increased radiative forcing by O_3 and decreased radiative forcing by CH_4 . OH resulting from LNO_x can also lead to less VSLs reaching the stratosphere.

7. Effect of Climate Change on Flash Rates

Lightning may be expected to increase as the climate warms because a warmer atmosphere holds more water according to the Clausius-Clapeyron relation and has higher values of CAPE (Füllekrug et al., 2022). However, a consensus does not exist (Clark et al., 2017; Murray, 2018; Williams, 2005) because warming will also affect the electric circuit and cloud microphysics in ways that may not yet be understood. In general, models using cloud-height- (Brasseur et al., 2006; Grenfell et al., 2003; Hauglustaine et al., 2005; Murazaki & Hess, 2006; Price & Rind 1994; Zeng et al., 2008) or CAPE-based flash rate schemes (e.g., Romps et al., 2014) show substantial increases in total flash numbers with warming, while schemes based on updraft strengths, ice flux or cold-cloud depth show only smaller increases or even decreases (Del Genio et al., 2007; Finney et al., 2018; Futyan & Del Genio, 2007). The increases in flash rates associated with PR-based schemes are not surprising, as flashes in that scheme vary with the 5th power of cloud top height, which has been shown to increase in most simulations of the future (e.g., Clark et al., 2017). Cloud-top heights have been shown to have an observed increasing trend (see below). First, we present some examples of long-term trend studies using lightning and cloud-brightness temperature data obtained from satellites and data from lightning detection networks. Second, we discuss the global model studies that include future projections of flash rates. The impact of future changes in global or regional flash rates on atmospheric chemistry is unknown. If individual storms have greater flash rates in the future, mean

Table 10
Observed Flash Rate Trends

Reference	Data source	Region	Flash rate change
Füllekrug et al. (2022)	LIS	Global	Unchanged 1998–2021, but –10% in 2020 and –8% in 2021
Liu et al. (2025)	GLD360 and WWLLN	Global	3%–6% reduction in 2020 relative to 2018–2019
Petersen and Buechler (2008)	LIS	Entire tropics and subtropics	Unchanged 1998–2007
Koshak et al. (2015)	NLDN	CONUS	–12.8% from 2003–2007 to 2008–2012
Qie et al. (2020)	LIS	Entire tropics and subtropics	Unchanged 1998–2013, but significant regional trends
Chakraborty et al. (2021)	LIS	India	+1% to +2.5% per year 1998–2014
Wang et al. (2021)	China Lightning Detection Network	Three urban areas	–50% to –75% 2010–2018
Pérez-Invernón (2021)	ENTLN	Po Valley, Italy	–89% in March/April 2020 from 2017 to 2019 mean
Holzworth et al. (2021)	WWLLN	>65°N	Fraction of global strokes north of 65°N increased by factor of 3 over 2010–2020
Vaisala (2022)	GLD360	>80°N	> an order of magnitude increase from 1st half to 2nd half of 2012–2021 period
Taszarek et al. (2021)	NLDN	CONUS	–5% to –15% over southern US in summer 1989–2018; lesser decrease in spring and autumn; mixed trends in winter
Qie et al. (2022)	OTD/LIS	Eastern Tibetan Plateau	+17% 1996–2013
Prasanna et al. (2024)	OTD/LIS	India	+32% premonsoon season; +34% monsoon season; 1996–2013
Burrows et al. (2025)	Canadian Lightning Detection Network	Canada	Decreases from 1999 to 2023 with greatest decrease over Ontario and Quebec. Increases found over northern territories.

PE will likely decrease as mean channel lengths will become smaller based on the findings discussed in Section 3. If there are a greater number of electrified storms in the future increasing the global flash rate, but the storm flash rates are such that global mean PE remains the same, the total LNO_x production will increase.

Observed flash rate trends are summarized in Table 10. Satellite-based (OTD-LIS) observations (Cecil et al., 2014) do not show a significant trend in global flash rates over the 1996 to 2013 time period although regional trends are evident, possibly due to regional trends in parameters such as CAPE or aerosol loading (e.g., Chakraborty et al., 2021; W. A. Petersen & Buechler, 2008; K. Qie et al., 2020; X. Qie et al., 2022). Füllekrug et al. (2022) examined trends in lightning over the 1998–2021 time period using a merged data set derived from the LIS sensors aboard TRMM and ISS. They found no trend with annual mean deviations less than 5% except for –10% in 2020 and –8% in 2021. They speculated that the negative deviations in 2020 and 2021 were due to decreases in aerosol amounts due to the Covid-19 lockdowns and general declines in economic activity. Y. Liu et al. (2025) linked 7.6% decreases in global aerosol optical depth (AOD) to 3%–6% flash reductions in 2020 relative to the 2018–2019 mean from the Vaisala Global Lightning Detection (GLD360) and World Wide Lightning Location (WWLLN) Networks. Other investigators have also noted significant flash rate reductions during the Covid-19 lockdowns. Pérez-Invernón et al. (2021) found large reductions in ground-based flash observations (particularly in March and April 2020) associated with the lockdown in the Po Valley of Italy. Approximately 40% of the flash reduction was attributed to reduced aerosol loading and the remainder due to changes in meteorology.

Other ground-based flash observational trends have been reported in the United States, China, India, and the Arctic. Koshak et al. (2015) reported a decrease of 12.8% in 5-year mean CG flashes observed by the National Lightning Detection Network over CONUS from 2003–2007 to 2008–2012. Decreases in both wet-bulb temperature and aerosol concentrations could have contributed to this decrease in flashes. H. Wang et al. (2021) found 50%–75% reductions in CG flashes recorded by the China Lightning Detection Network over three polluted urban areas during the period of 2010–2018. Aerosol Optical Depth also decreased over these cities during this time period. Prasanna et al. (2024) found an increasing trend in lightning over India during 1996–2013 using the

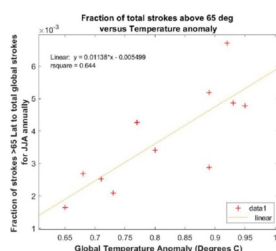


Figure 26. Direct comparison of the fraction of WLLN strokes above 65°N to total WLLN strokes (vertical axis), and the global temperature anomaly in degrees C (horizontal axis). From Holzworth et al. (2021).

regions north of 60°N latitude by the 2090s. Tazarek et al. (2021) examined long-term trends (1979–2019) in frequency of environments conducive for atmospheric convection using hourly ERA-5 reanalysis data. They found decreases in instability and increased convective inhibition over the southern US in summer. Positive trends were found in south-central and northern Europe. Analysis of NLDN CG flash data also showed decreases of 5%–15% over the southern US in summer during 1989–2018, lesser decreases in spring and autumn, and mixed trends in winter.

Richardson et al. (2022) examined trends in the height of tropical upper tropospheric clouds over the 2002 to 2021 time period using retrievals from Moderate Resolution spectroradiometer (MODIS) instruments aboard the Terra and Aqua satellites. They observed increases of $6.9 \pm 2.7 \text{ m yr}^{-1}$ that were statistically consistent with modeled increases from global models. Of course, regional trends can be larger. Antony et al. (2025) investigated changes in the brightness temperature of deep convective clouds over the Indian subcontinent and adjacent oceanic region during the monsoon season over the 2000 to 2020 time period. They found that the decreases in brightness temperature were associated with increases of $\sim 1 \text{ km}$ in the heights of deep convective clouds.

He and Sudo (2023) explored lightning trends over the 1960–2014 time period using the CHASER model with the cloud-height and ECMWF-McCAUL flash rate schemes. Globally, they found that increases in flash rates due to global warming ($3\% \text{ K}^{-1}$) were balanced by decreases in flash rates due to aerosol radiative effects associated with increases in aerosols and their precursors. Specifically, He and Sudo (2023) noted that aerosol radiative effects weakened convective activity over lightning hotspots. They cautioned that their model simulation did not include sub-grid scale aerosol microphysical effects, which are believed to increase flash rates via convective invigoration. He and Sudo (2023) also state that further investigation is needed due to the contrast between their flat trends obtained using prescribed sea-surface temperatures and the positive trends of three CMIP6 models that used a coupled atmosphere-ocean GCM to calculate SSTs.

Model projections of future flash rates have produced a variety of results. Such studies published since 2007 are summarized in Table 11. Clark et al. (2017) examined current year and future year (RCP 4.5 and RCP 8.5) flash rates from multiple schemes in CAM5, the atmospheric component of the CESM. They found that a cloud-height based scheme predicted a $\sim 12\%$ increase in flash rate per degree of warming while a cold-cloud-depth based scheme predicted a 3.7% increase. Interestingly, convective-precipitation-based and mass flux based schemes predicted increases in flash density for the RCP4.5 case but decreases for the RCP8.5 case. Finney, Doherty, Wild, Young, and Butler (2016b) examined the response of LNO_x to climate change in 12 models that participated in the Atmospheric Chemistry and Climate Model Intercomparison Project (ACCMIP). Ten of these models used the cloud-top height flash rate parameterization and showed an average response of $+0.44 \pm 0.05 \text{ Tg N K}^{-1}$ after normalizing the sources to 5 Tg N yr^{-1} . One model used the Allen and Pickering (2002) updraft cloud mass flux at 440 hPa scheme and had a response of $-0.55 \text{ Tg N K}^{-1}$, while a final model which used the Grewe et al. (2001) combined updraft mass flux and cloud-top height scheme showed a response of $+0.14 \text{ Tg N K}^{-1}$. Thornhill et al. (2021) examined the response of flash rates in four Earth System Models with interactive chemistry to a doubling of CO_2 . While each of the models parameterized flash rates in terms of cloud top height, the change in lightning- NO_x production per 1° increase in temperature ranged from -0.03 to $0.61 \text{ Tg N per year}$.

Finney et al. (2018) found a 15% decrease in flash rates for a 4.87 K warming under an RCP 8.5 scenario, when parameterizing flash rates in terms of upward cloud ice flux. Roms (2019) examined the sensitivity of flash rates to temperature using a 4-km resolution cloud-resolving model over the CONUS with four proxies for flash rate.

Table 11
Projected Future Flash Rate Trends

Reference	Flash rate parameterization	Flash rate change
Romps et al. (2014)	CAPE $\times$ Precip Rate	CG only +12.5% per $^{\circ}\text{C}$
Finney et al. (2016b)	Cloud-top height; convective mass flux; Grewe et al. (2001)	+8.8%; -11% ; +2.8% per $^{\circ}\text{C}$ by 2100 in ACCMIP simulations
Clark et al. (2017)	Cloud-top height; cold-cloud depth; convective mass flux	+12%; +3/7%; -2% per $^{\circ}\text{C}$ for RCP8.5
Finney et al. (2018)	Upward ice flux	-15% by 2100 with RCP8.5 warming of 4.87°C
Romps (2019)	CAPE $\times$ Precip Rate; Precip Rate with vertical velocity $> 10\text{ m/s}$; upward ice flux at 260 K; Cloud ice $\times$ graupel	CONUS: +8% to +16% per $^{\circ}\text{C}$ by 2071–2100; Tropics: CAPE and precip schemes: +5% to +12% per $^{\circ}\text{C}$; ice schemes: -1% to -4% per $^{\circ}\text{C}$
Thornhill et al. (2021)	Cloud-top height in four models	+5.4% to +12.2% per $^{\circ}\text{C}$ for $4\times\text{CO}_2$
Chen et al. (2021)	CAPE $\times$ Precip Rate	+112% in boreal permafrost regions with RCP8.5
Charn and Parishani (2021)	CAPE $\times$ Precip Rate; McCaul et al. (2009); Upward ice flux; IWC $\times$ graupel	0% to +6% per $^{\circ}\text{C}$ for 4°C warming
Kahraman et al. (2022)	McCaul et al. (2009)	Decreases over most of Europe by 2100 under RCP8.5
He et al. (2022)	Upward ice flux; cloud-top height; ECMWF-Lopez (2016); ECMWF-McCaul	-4% to -9% ; +2.5%; +13%; +10% per $^{\circ}\text{C}$ over 2001–2020
Michibata (2024)	ECMWF-McCaul	+18.4% per $^{\circ}\text{C}$ with SST $+4^{\circ}\text{C}$
Whaley et al. (2024)	Etten-Bohm et al. (2021)	Globally no change by 2081–2100; +6% northern mid-lat; -8% tropics per $^{\circ}\text{C}$ with SSP5-8.5
Etten-Bohm et al. (2024)	Etten-Bohm et al. (2021)	Globally -5% ; high lat +50% by 2098–2100 with RCP8.5

Estimates of flash rates in 2071–2100 were made assuming the warming rate in the CMIP5 model ensemble (Figure 27). CONUS flashes increased 8%–16% per $^{\circ}\text{C}$. Tropical marine flash rates sensitivity to temperature were assessed in a radiative-convective equilibrium model. Flash rates were found to increase 5%–12% per $^{\circ}\text{C}$ with the precipitation-based schemes, but decrease 1%–4% per $^{\circ}\text{C}$ with the ice-based schemes.

Charn and Parishani (2021) examined the response of several lightning proxies to a 4°C warming using a cloud-resolving model with 1-moment and 2-moment microphysics schemes embedded within CAM5. The proxies were CAPE $\times$ PRETOT (total precipitation rate) (Romps et al., 2014), graupel flux (McCaul et al., 2009), ice flux across the 258 K isotherm (Finney et al., 2014; Romps, 2019), and ice water content (IWC) $\times$ graupel. They found that the simulations with 2-moment microphysics did the best job of reproducing the climatology of present-day flashes. When 2-moment microphysics were invoked, the response of flash rates in the 5 schemes to warming was broadly similar, that is, increases of 0%–6% K^{-1} depending on details of the parameterization schemes.

Per 1°C increase in global mean surface temperature, He et al. (2022) found that LNO_x production decreased by 4%–9% for an ICEFLUX-based scheme, increased by 2.5% for a cloud-height scheme, increased by 13% for the ECMWF Lopez (2016) scheme, and increased by 10% for an ECMWF version of the McCaul et al. (2009) scheme in a CHIMERE (MIROC) simulation that spanned 2001–2020. Michibata (2024) added a 2-moment prognostic graupel scheme to the MIROC6 GCM. The model performed well in reproducing distributions of graupel, CAPE, and lightning flash rates compared with reanalysis and satellite data. With future warming specified as a uniform $+4^{\circ}\text{C}$ SST increase, the global flash rate increased 18.4% per $^{\circ}\text{C}$ using the ECMWF-McCaul scheme.

Recently, Whaley et al. (2024) ran 10-member ensembles of CanESM5.1 with the Etten-Bohm et al. (2021) lightning scheme into the future (years 2081–2100) under climate change scenarios, SSP5-8.5 and SSP2-4.5. Under the first higher-emissions scenario, CanESM5.1 predicted almost no change to the global mean lightning flash rate with flash rates increasing in the northern midlatitudes by $\sim 6\%$ but decreasing in the tropics by $\sim 8\%$. Using the Etten-Bohm et al. scheme, Etten-Bohm et al. (2024) examined changes in global flash rates between

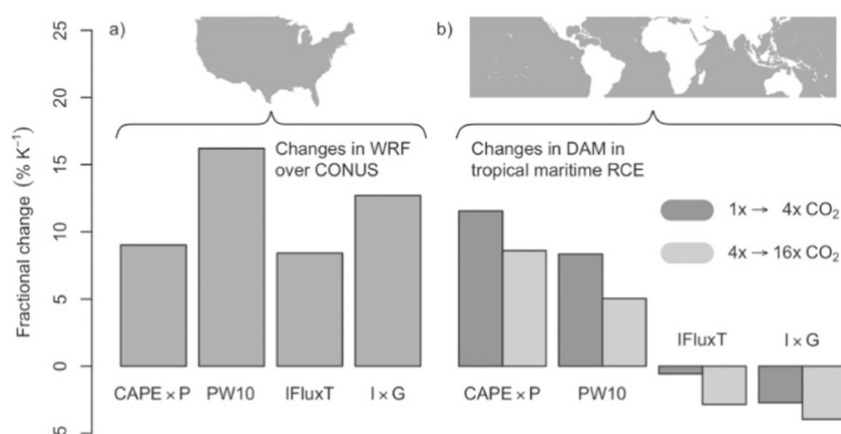


Figure 27. Fractional change in continental US (a) and tropical marine (b) flash rates per degree change in surface temperature (a) or sea surface temperature (b). Continental U.S. changes obtained from WRF-simulations nudged to represent present-day and future-conditions. Tropical marine changes obtained using DAM Radiative Convective Equilibrium (RCE) model simulations with present-day and 4 × present day CO₂ concentrations. The flash rate proxies used were CAPE × Precipitation Rate; Precipitation Rate with vertical velocity > 10 m s⁻¹; upward ice flux at 260 K isotherm; and cloud ice × graupel. From Roms (2019).

2017–2019 and 2098–2100 in CAM5 under a RCP8.5 climate scenario. Despite high-latitude increases, of ~50%, the global mean lightning occurrence was predicted to decrease by about 5%.

Observations of flash rate from satellite have revealed no significant trends and typically no more than 5% interannual variation. The exception to this finding is a ~10% reduction in flashes in 2020 and 2021 that has been tied to reduction in aerosol loading during the Covid-19 pandemic. Numerous positive and negative regional trends in flashes have been detected by satellite and ground-based observation systems. Future lightning trends have been estimated in global models using several different flash rate parameterization schemes. Typically, the commonly used cloud-top height scheme predicts significantly greater flash rates under future warming scenarios, while the upward ice flux scheme predicts future decreases in flash rates. Other schemes produce either negative or mixed results. In summary, there remains considerable uncertainty concerning future flash rates, and observations to date do not suggest any long-term global trend despite the warming that has already occurred. Significant changes in flash rates that have been observed regionally appear to be related to changes in atmospheric aerosol content.

Extratropical forests store more carbon per unit area than tropical forests. Permafrost soils contain even larger reservoirs of carbon (Chen et al., 2021). This carbon is released in the form of CO₂, CO, and aerosol when fires occur. Lightning is estimated to account for 77% of the burned area in extratropical fires (Janssen et al., 2023) with long-current flashes (LCC) accounting for a sizable portion of that percentage (Pérez-Invernón, Gordillo-Vázquez, Huntrieser, & Jöckel, 2023; Pérez-Invernón, Gordillo-Vázquez, van der Velde, et al., 2023). Extratropical flash rates are predicted to increase by 10%–30% per degree warming with the lower bound obtained from an ice-flux based scheme and the upper bound from a cloud-height based scheme. These increases may already be occurring (e.g., Holzworth et al., 2021; Vaisala, 2022; see 3rd paragraph of this section).

Lightning-initiated fires over regions covered by permafrost could increase by more than the extratropical mean. Utilizing CMIP-5 models under the RCP8.5 high-emissions scenarios and parameterizing flash rates in terms of the product of CAPE and precipitation, Chen et al. (2021) found that summertime lightning over regions currently underlain by permafrost increased by $112 \pm 38\%$ by the end of the century. They noted that these changes would make flash rates at the northern tree line comparable to current flash rates 480 km to the south. The percent increase in burned area could be larger if the increase in lightning is accompanied by vegetation feedbacks that accelerate the northern migration of trees and the release of carbon from thawing permafrost (see also Finney, 2021; Natali et al., 2021). The percent increase could be smaller if climate-change related increases in RH and precipitation reduce the chance that any one flash creates a fire (Pérez-Invernón, Gordillo-Vázquez, Huntrieser, & Jöckel, 2023). Regardless, given the large amount of carbon stored in permafrost, it is essential to

increase monitoring of high latitude fires and to enhance coupling between vegetation models and climate models that include lightning and LCC lightning parameterizations.

8. Summary and Conclusions

The following bullets summarize the conclusions of this literature review (2007–2025) concerning the effects of lightning on atmospheric chemistry:

- Based on over 40 estimates of LNO_x PE published between 2007 and 2025 we find that the most likely range for global LNO_x production is $5 \pm 2 \text{ Tg N yr}^{-1}$, which represents a small reduction in the width of the range from that published by Schumann and Huntrieser (2007) ($5 \pm 3 \text{ Tg N yr}^{-1}$).
- Observational (satellite and aircraft) and theoretical/laboratory-based estimates of PE tend to be smaller ($145\text{--}192 \text{ mol fl}^{-1}$; $2.9\text{--}3.8 \text{ Tg N yr}^{-1}$) than those based on global, regional, and cloud-resolved models that are constrained by observations ($355\text{--}374 \text{ mol fl}^{-1}$; $7.1\text{--}7.5 \text{ Tg N yr}^{-1}$).
- LNO_x PE per flash has been found to be inversely related to flash rate by several authors. This finding is consistent with data from LMAs showing an inverse relation between flash size and flash rate. This characteristic of the LNO_x source is now being incorporated in chemical transport and climate models.
- Microphysically based flash rate schemes produce the most realistic distribution of flashes, provided ice microphysics is adequately treated in a model.
- The ratio of PE_{CG} to PE_{IC} still remains one of the largest uncertainties in parameterizing LNO_x in models. There is evidence for a ratio near 1 and for a ratio much greater ($\sim 8\text{--}16$).
- Estimates of O_3 production in corona discharges have become larger in recent years, and this O_3 source may be significant enough to be included in chemical transport models. N_2O has also been found to be produced in corona discharges, but only in small quantities.
- LHO_x , largely produced by corona and subvisible discharges, is likely to play a non-negligible role ($2\%\text{--}16\%$) in the HO_x budget and should be included in chemical models.
- LNO_x leads to photochemical O_3 production in the UT at rates of $5\text{--}10 \text{ ppbv d}^{-1}$. Larger rates have been observed in the near field of deep convection, perhaps due to availability of freshly transported VOCs providing radical species.
- At most midlatitude locations, LNO_x contributes $1\text{--}2 \text{ ppbv}$ to maximum 8-hr surface O_3 during the lightning season; however, contributions of $\sim 6 \text{ ppbv}$ are not uncommon over higher terrain, where high boundary layer depths allow LNO_x -induced photochemically produced O_3 to mix to the surface.
- Multi-model estimates of the sensitivity of CH_4 lifetime to LNO_x is $\sim 5\%$ per Tg N yr^{-1} change of LNO_x production.
- Globally, no significant trends in lightning flash rate have been observed, except for a $\sim 10\%$ reduction in flashes during the 2020–2021 COVID-19 pandemic, which has been tied to reduction in aerosol loading. Both positive and negative regional trends have been detected by satellite and ground-based networks.
- CAPE and Cloud-top height-based parameterization schemes predict greater future flash rates under climate warming scenarios. Microphysical and other schemes yield either negative changes or mixed results for future flash rates. As a result, future flash rate trends and LNO_x emissions remain uncertain.
- Geostationary satellite observations of both the diurnal variation of lightning flashes and NO_2 columns that are now available will allow more detailed analysis of LNO_x PE and further reduction of its uncertainties.
- An important key to further reducing the uncertainty in LNO_x PE is to better understand the reasons for the distinct difference between theoretical/laboratory, aircraft and satellite mean PE values ($\sim 3 \text{ Tg N yr}^{-1}$) and those obtained from cloud, regional and global models ($\sim 7 \text{ Tg N yr}^{-1}$).

Conflict of Interest

The authors declare no conflicts of interest relevant to this study.

Availability Statement

Data were not used, nor created for this research.

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