Analysis of new observational datasets relating to the organization and dynamical impacts of tropical convection

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Close-up view of the text, which appears to be a section of a research paper or dissertation. The text discusses observational datasets relating to tropical convection, particularly focusing on the interplay between sea breezes, mountain-valley wind regimes, and remotely forced gravity waves. It also examines the relationship between lightning and nitrogen oxide radicals, with observations from WWLLN and GOME-2 satellites. The text is well-structured and clearly outlines the research focus and findings.
Lightning and NO$_2$ also vary coherently with the MJO, with variations of up to ~50% of the annual mean.

MJO-related deep convection induces planetary-scale Kelvin and Rossby waves in the stably stratified tropical tropopause transition layer (TTL). The structure of these waves is investigated using satellite observations from COSMIC, CALIPSO, and MLS, as well as ERA-Interim wind and humidity fields. Regions of ascent in the planetary waves are associated with anomalously low temperatures, high radiative heating rates, enhanced cirrus occurrence, and high carbon monoxide and low ozone concentrations. Low water vapor concentrations lag the low temperature anomalies by ~1-2 weeks. Anomalies in each field tilt eastward with height in the TTL and propagate downward from the lower stratosphere to the upper troposphere. As the Kelvin wave front propagates eastward across equatorial South America and Africa, equatorially-symmetric, anomalously low temperature and water vapor mixing ratio and enhanced TTL cirrus are observed above ~100 hPa in the zonal-mean.
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Dedication

To my parents, David and Shirley Virts, and my grandparents, Donald and Ethel Virts and August and Hermine Kettelhake, with loving thanks for their legacy of faith and hard work.
**Introduction**

Tropical weather and climate have long been a subject of interest to atmospheric scientists due in part to their role in the general circulation of the atmosphere. Weather in the tropics is modulated by phenomena which are in many ways distinct from the synoptic-scale disturbances that habitually produce day-to-day weather variations in mid-latitudes. Prolonged records of atmospheric variability have historically been scarce in the tropics, particularly over its extensive oceanic regions and numerous developing countries. Surface weather stations and radiosonde observations have long provided a means of investigating the tropical climate and the vertical structure of the tropical atmosphere. Analysis of radiosonde data led to the discovery, or permitted a refining of the scientific understanding, of such phenomena as the periodic shifts in zonal winds in the tropical stratosphere known as the Quasi-Biennial Oscillation (Reed 1966), Yanai (Yanai and Maruyama 1966) and Kelvin waves in the lower stratosphere (Wallace and Kousky 1968), the seasonal cycle in the temperatures of the TTL and lower tropical stratosphere (Reed and Vlcek 1969), and the intraseasonal Madden-Julian Oscillation (Madden and Julian 1971, 1972). In addition to the climate records available from weather observing stations, a series of field campaigns, including GATE and TOGA COARE, have provided intensive observations of the characteristics of tropical convection. GATE emphasized cloud structure and the organization of tropical convection at spatial scales ranging from individual clouds to mesoscale clusters (Houze and Betts 1981, and references therein), while TOGA COARE produced observations of the evolution of mesoscale convection systems (Rickenbach and Rutledge 1998) and their relationship to the MJO (Chen et al. 1996), as well as the diurnal variability of convection (Sui et al. 1997).
Satellites have, over the last few decades, increasingly provided alternative observations to complement the local station or short-term field campaign datasets. Both regular measurement of basic atmospheric variables such as temperature in previously poorly-sampled regions (Spencer and Christy 1993) and observation of trace gas atmospheric constituents such as ozone (McCormick et al. 1989) are examples of the developments made possible by satellites. In the course of the last 5-15 years, a series of satellites have been launched with the goals of determining the distribution and variability of precipitation (TRMM), detecting trace gas concentrations (GOME-2 and MLS), and providing high-resolution vertical profiles of temperature (COSMIC) and cloud and aerosol layers (CALIPSO). The availability of these satellite datasets represents a unique new opportunity to investigate existing research questions regarding tropical weather and climate.

It will become clear when reading this thesis that a common theme is analysis of various atmospheric fields with respect to the MJO. Attaining a better understanding of the MJO itself is, however, only a secondary goal of this thesis. Rather, the MJO is used as a carrier signal. The basic physical and dynamical relationships among temperature, circulation, clouds, and trace gases in the TTL are investigated by analyzing MJO-related perturbations based on satellite observations and reanalysis fields (Chapter 1). The role of lightning in \( \text{NO}_x \) production is confirmed by comparing the evolution of satellite-observed \( \text{NO}_2 \) and ground-based lightning observations during the MJO (Chapter 2). The importance of local, diurnally-varying circulations in driving thunderstorm occurrence over the Maritime Continent is demonstrated by analyzing, and predicting, lightning anomalies during the MJO, which is associated with large-scale wind and cloudiness anomalies that modulate the strength of the local circulation regimes (Chapter 4).
Many of the satellite datasets analyzed in this thesis have high vertical and/or along-track horizontal resolution. CALIPSO, for example, produces very detailed cross-sections of cloud and aerosol layers beneath the satellite’s path. Before statistical analysis can be performed, however, it is necessary to produce gridded datasets based on the satellite observations. Complicating this process is the fact that consecutive daytime or nighttime equatorial crossings of the polar-orbiting satellites are ~25° longitude apart. CALIPSO views only a so-called “curtain” of the atmosphere below the satellite, and the COSMIC profiles for any given day are distributed unevenly around the globe. Because of these characteristics, it is not possible to generate meaningful daily maps of cloudiness, temperature, or trace gases based on the CALIPSO, COSMIC, or MLS data. These limitations must be taken into account when choosing time averaging intervals and grid spacing for the gridded datasets and when choosing which research problems to address using a given dataset. The satellite datasets analyzed in Chapter 1 have been gridded rather coarsely, but weekly-averaged data on a 10° latitude × 10° longitude grid is sufficient to illustrate the planetary-wave features associated with the MJO. GOME-2, on the other hand, views a wide swath of the Earth, so that maps combining all the observations for a given day have only narrow regions of missing data. Ground-based WWLLN provides continuous, global sampling. As a result, it is possible to investigate day-to-day variations in lightning NOx production (Chapter 2) and to generate and analyze high-resolution, hourly climatologies of lightning observations (Chapters 3 and 4).
Chapter One: Observations of temperature, wind, cirrus, and trace gases in the tropical tropopause transition layer during the MJO

(Based on the following publication: Virts, K. S., and J. M. Wallace, in preparation)

Abstract

Satellite observations of temperature, cirrus, and trace gases derived from the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC), Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO), and Microwave Limb Sounder (MLS) are analyzed in combination with European Centre for Medium-Range Weather Forecasts (ECMWF) Re-analysis (ERA)-Interim wind and humidity fields in the tropical tropopause transition layer (TTL) using the Madden-Julian Oscillation (MJO) as a carrier signal. MJO deep convection induces planetary-scale Kelvin and Rossby waves in the stably stratified TTL. Regions of ascent in these waves are associated with anomalously low temperatures, high radiative heating rates, enhanced cirrus occurrence, and high carbon monoxide and low ozone concentrations. Low water vapor mixing ratio anomalies lag the low temperature anomalies by ~1-2 weeks. Anomalies in each field are observed to propagate eastward, circumnavigating the tropical belt over an interval of ~40 days. Equatorial cross-sections reveal that the anomalies tilt eastward with height in the TTL and propagate downward from the lower stratosphere to the upper troposphere.

As MJO convection moves into the western Pacific and dissipates, a fast-moving Kelvin wave front propagates eastward across equatorial South America and Africa into the western Indian Ocean. The region of westerly wind anomalies behind the front lengthens until it encompasses most of the tropics at the 150-hPa level, with equatorially-symmetric, anomalously
low temperature and water vapor mixing ratio and enhanced TTL cirrus above ~100 hPa in the zonal-mean.

Section 1.1. Introduction

In the tropics, the troposphere and stratosphere are linked by the tropical tropopause transition layer (TTL). The TTL lies above the tops of most deep convective clouds (Alcala and Dessler 2002) and above the altitude of zero net radiative heating (Corti et al. 2005). The climatological-mean TTL is characterized by mean upward mass flux that links the tropospheric Hadley circulation with the stratospheric Brewer-Dobson circulation (Corti et al. 2005), and the upper limit of the TTL can be described as the level where the upward mass flux matches that in the lower stratospheric Brewer-Dobson circulation (Fu et al. 2007). Temperatures near the tropical cold point, around 90 hPa, are lower than those observed at similar altitudes in the mid-latitudes and those at the mid-latitude tropopause (Fueglistaler et al. 2009a). The lowest temperatures and the greatest prevalence of TTL cirrus clouds tend to be observed over the areas of frequent deep convection over the equatorial landmasses of Africa, South America, and the Maritime Continent (Fueglistaler et al. 2009a; Virts and Wallace 2010). The TTL is modulated on a variety of temporal scales by stratospheric phenomena, such as the annual cycle in the strength of the upwelling in the Brewer-Dobson Circulation (Yulaeva et al. 1994; Dima and Wallace 2007) and the descending zonal wind regimes in the Quasi-Biennial Oscillation (Randel et al. 2000), and by phenomena associated with the tropical troposphere and oceans, such as the interannual El Niño/Southern Oscillation (Randel et al. 2000; Gettelman et al. 2001; Kiladis et al. 2001).
Tropical atmospheric variability at intraseasonal time scales (a ~30-80 day cycle) is dominated by the Madden-Julian Oscillation (MJO; Madden and Julian 1971, 1972). An active period of the MJO begins with the development of a region of enhanced convection with low-level convergence over the equatorial Indian Ocean. The convective envelope propagates eastward across the Maritime Continent into the western Pacific warm pool and dissipates in the central Pacific (Madden and Julian 1994; Zhang 2005). The tropospheric heating produced by the MJO convection induces planetary-scale perturbations in the TTL that are not confined to the equatorial Indian and Pacific regions but rather propagate throughout the tropical belt (Hendon and Salby 1994).

In recent years, a series of satellites, described in Section 1.2, have provided unprecedented sampling of temperature, clouds, and trace gas concentrations in the TTL. In this study, we examine these atmospheric fields in relation to each other, using the MJO as a carrier signal. We first review the basic structure of the MJO-induced planetary wave pattern and then investigate the relationships among the MJO-related anomalies at a single level (100 hPa) in the TTL and in longitude-height cross-sections above the equator, using both reanalysis fields and a variety of satellite observations (Section 1.3). Analysis of zonally-averaged MJO perturbations in the TTL is in Section 1.4, and conclusions are in Section 1.5.

Section 1.2. Data and analysis techniques

a. Data sources

The six satellites in the Constellation Observing System for Meteorology, Ionosphere, and Climate (COSMIC)/Formosa Satellite 3 (FORMOSAT-3) measure the atmosphere’s refraction of radio waves from GPS satellites (Anthes et al. 2008). On average, over 1000
tropical temperature profiles per day are obtained from these radio occultations. COSMIC temperature profiles extend from 40 km down nearly to the Earth’s surface with 100-m vertical resolution. NCAR has made available COSMIC profiles corrected for the effects of water vapor on the GPS signal. We make use of these corrected profiles even though the water vapor effect is small in the relatively dry and cold TTL and lower stratosphere (Kurskin斯基 et al. 1996; Anthes et al. 2008).

The Cloud-Aerosol Lidar and Infrared Pathfinder Satellite Observations (CALIPSO) and Aura are polar-orbiting satellites in NASA’s Afternoon constellation (or “A-train”); Aura flies just minutes behind CALIPSO in the constellation. CALIPSO carries a two-wavelength polarization lidar that is capable of detecting cloud layers with optical depths of 0.01 or less. Cloud layer data are available at 5 km along-track resolution and 60 m vertical resolution for the altitudes of interest (Winker et al. 2007). The lidar signal can become completely attenuated in optically thick clouds (Winker et al. 2007); following Fu et al. (2007), we assign such opaque layers a cloud base at the earth’s surface.

The Microwave Limb Sounder (MLS) aboard the Aura satellite measures thermal emissions in spectral bands centered on 190 and 240 GHz. From these radiance measurements are derived vertical profiles of ozone, carbon monoxide, and water vapor mixing ratio at 1.5° intervals along the satellite’s orbital path (Read et al. 2007; Livesey et al. 2008). In this study, we analyze MLS ozone and carbon monoxide mixing ratios at 100 hPa and water vapor mixing ratios at levels between 316 hPa and 46 hPa. The vertical resolution of the MLS water vapor profiles varies with height and is ~3 km near 100 hPa (Read et al. 2007); ozone and carbon monoxide profiles have vertical resolutions of ~2.5 km and ~4 km in the TTL, respectively (Livesey et al. 2008).
Fields of other atmospheric variables are represented in this study by the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA)-Interim (Dee et al. 2011). The 100-hPa ERA-Interim analyses examined in Section 1.3 are available four times daily at 1.5° horizontal resolution, spanning the globe. In addition, we examine daily ERA-Interim zonal and meridional winds, as well as shortwave and longwave radiative heating rates, at 1° horizontal resolution and at 14 vertical levels between 320 and 44 hPa. Adding together the shortwave and longwave heating rates produces net radiative hating rates, which are referred to as “heating rates” in this study. Ascent in the TTL and lower stratosphere produces adiabatic cooling that is balanced by radiative heating, as the air layer undergoes radiative relaxation to the ambient temperature (Andrews et al. 1987; Fueglistaler et al. 2009b). We will investigate how radiative heating rates compare with other means of inferring vertical velocity in the TTL.

This study also makes use of NOAA’s outgoing longwave radiation (OLR) observations, which are available at 2.5° latitude × 2.5° longitude resolution.

b. Three-dimensional indices

In this study, we analyze four years (13 June 2006 to 12 June 2010) of observations from each of the satellites listed above and ~3.5 years (1 April 2006 to 31 December 2009) of ERA-Interim data. As in Virts et al. (2010), a weekly TTL cirrus index is calculated on a 10° latitude × 10° longitude grid and is defined as the fraction of CALIPSO cloud profiles acquired within each grid box that identified a cloud layer with a base above 15 km. For each 10° × 10° grid box, a weekly height-dependent cloud index is also calculated as the cloud fraction within successive 200-m layers. Similarly, a zonal-mean height-dependent cloud index is calculated as cloud fractions within 5° latitude bands. These indices are referred to as “cloud fraction” in this
paper. Similar spatial and temporal averaging is performed on ERA-Interim analyses, MLS trace gas observations, and COSMIC temperature profiles at each vertical level. On average, CALIPSO and MLS sample some portion of a $10^\circ \times 10^\circ$ grid box in the equatorial belt five times per week, and COSMIC averages $\sim 15$ temperature profiles per week per grid box.

After spatially and temporally averaging the data, an 80-day high-pass Lanczos filter is applied to the time series at each grid box. The Student’s $t$ test is used to determine the statistical significance of correlation coefficients. The filtered time series analyzed in Section 1.3 have an average of $\sim 166$ effective degrees of freedom, estimated using the formula of Leith (1973), so if a two-sided distribution is assumed, correlations of 0.16 in absolute value are statistically significant at the 95% level. Zonal-mean fields average $\sim 77$ degrees of freedom, so that correlations of 0.22 are significant at the 95% level.

### Section 1.3. MJO signature in 100-hPa maps and equatorial cross-sections

#### a. Planetary-wave signature

To quantify the evolution and strength of the MJO, we use the Real-time Multivariate MJO (RMM) index introduced in Wheeler and Hendon (2004). The two components of this index, RMM1 and RMM2, are the time series produced by projecting daily observations, with annual and interannual variability removed, onto principal component time series associated with the first two empirical orthogonal functions (EOFs) of a field of daily, near-equatorial OLR and zonal wind at 850 and 200 hPa. RMM1 and RMM2 are used to define the phase space illustrated in Fig. 1.1, which has been modified from Wheeler and Hendon (2004). An idealized evolution of the MJO appears as a counterclockwise rotation when represented on this chart.
Figure 1.2 shows OLR patterns obtained by regressing 80-day high-pass filtered OLR time series onto RMM1 and RMM2; for clarity, we also show OLR patterns associated with the linear combinations RMM1-RMM2 and RMM1+RMM2, which fall one-eighth cycle before RMM1 and RMM2, respectively. A broad area of enhanced cloudiness, indicated by anomalously low cloud-top temperatures, develops over the equatorial Indian Ocean and propagates eastward over the Maritime Continent into the western and central Pacific Ocean, followed by an area of suppressed cloudiness. Weak OLR anomalies are observed over Africa and the Americas. By construction, the remaining half of an idealized MJO cycle is associated with OLR patterns with signs opposite of those shown in Fig. 1.2. Thus, as indicated in Fig. 1.1, RMM1 can be thought of as an MJO-related pulsation of cloudiness and precipitation over the Maritime Continent and RMM2 as a dipole with opposing centers in the Indian Ocean and the western Pacific Ocean.

The 100-hPa temperature anomalies associated with the MJO, based on COSMIC GPS radio occultation temperature profiles, are shown in Fig. 1.3. Similar patterns and amplitudes are obtained by analyzing COSMIC cold-point temperatures or ERA-Interim 100-hPa temperatures (not shown). Cross-sections of near-equatorial (5°S to 5°N) COSMIC temperature anomalies for the same MJO phases are shown in Fig. 1.4. The MJO temperature perturbations shown in Fig. 1.3 and 1.4 exhibit the following characteristics:

- At 100 hPa, a planetary-scale region of anomalously low temperatures propagates eastward along the equator ~30° of longitude to the east of the enhanced MJO convection; likewise, higher temperatures are observed to the east of the areas of suppressed MJO convection. Anomalously low 100-hPa temperatures and high
tropopause heights above and to the east of the convection were also observed by Madden and Julian (1972), based on radiosonde data.

- The temperature field above the warm pool region resembles the modeled response to a heat source situated on the equator (Matsuno 1966; Webster 1972; Gill 1980). Low temperatures extend along the equator to the east of the temperature minimum and tilt eastward with height (Fig. 1.4), a configuration that is consistent with the downward wave propagation and upward energy dispersion in an equatorially-trapped Kelvin wave (Holton and Lindzen 1968; Wallace and Kousky 1968; Holton 1979). Subtropical cold anomalies associated with the Rossby waves are observed poleward of the enhanced convection and extend westward for over 90° of longitude (Fig. 1.3).

- As enhanced MJO convection moves from the Maritime Continent into the western and central Pacific, a fast-moving Kelvin wave front propagates rapidly eastward from the central Pacific, resulting in low temperatures over equatorial South America and Africa at the time of maximum RMM2 (Fig. 1.3d, 1.4d). Subtropical warm anomalies flank the cold anomalies behind the Kelvin wave front.

Aspects of the structure and evolution of the MJO-related planetary-scale temperature perturbations in the upper troposphere and TTL have been previously documented based on temperature observations from the Microwave Sounding Unit satellite (Hendon and Salby 1994; Bantzer and Wallace 1996), GPS radio occultation measurements including COSMIC (Tian et al. 2012), and reanalysis datasets (Kiladis et al. 2001; Dima and Wallace 2007).

The results shown thus far have illustrated the location and basic structure of the equatorial planetary waves associated with the MJO. In the remainder of this section, we
emphasize the coherent picture that emerges by examining MJO-related perturbations based on data from diverse satellite and reanalysis sources.

b. Multi-variate analysis

In this section, we examine the anomaly patterns of circulation, clouds, and trace gases in the TTL produced by the planetary waves induced by the MJO deep convection. Due to space limitations, only fields associated with RMM2 are shown. To facilitate comparison among the various fields, the RMM2 temperature patterns from Fig. 1.3 and 1.4 are repeated in contours in each panel of all subsequent figures and are overlaid with colored shading representing other atmospheric fields. For each field shown, the on-line supplementary material includes an animation showing its evolution throughout the MJO cycle. We will occasionally refer to these animations in the text, and we encourage readers to view them to gain a fuller understanding of the MJO’s impact on the TTL.

MJO-related zonal and meridional wind and vertical velocity anomalies from the ERA-Interim 100-hPa analyses are shown in Fig. 1.5a, b. The 100-hPa winds converge into the region of anomalously high temperature above the Maritime Continent and diverge from the region of anomalously low temperature over the equatorial central Pacific. Westerly winds are observed from the equator out to ~20° latitude over South America and Africa, in association with what appear to be equatorial Kelvin waves and cyclonic Rossby waves in the subtropics. Anomalies of vertical velocity in pressure coordinates from ERA-Interim, shown in Fig. 1.5b, indicate subsidence (i.e., positive values) at 100 hPa above the western Pacific warm pool associated with the warm phase of the Kelvin wave; however, the remainder of the ERA-Interim vertical velocity field exhibits little or no resemblance to the overlaid temperature field. The RMM2 96-hPa
heating rate anomaly pattern shown in Fig. 1.5c exhibits much closer spatial agreement with the pattern of temperature anomalies—diabatic heating in association with cold anomalies, and vice versa. A strong correspondence is also observed between diabatic heating and independently detected fields such as TTL cirrus and trace gases, as will be shown later.

An equatorial cross-section of RMM2 circulation anomalies, as represented by zonal wind and heating rates, is shown in Fig. 1.6. The heating rates are represented as vertical vectors, scaled by multiplying their correlation coefficient with RMM2 by a factor of -1, so that upward-point vectors indicate positive heating rates. The circulation and temperature anomalies illustrate the anomalous warmth and divergence in the upper troposphere, around the 250-hPa level, associated with the enhanced convection over the western and central Pacific, just east of the Date Line (Fig. 1.2d). Westerly winds and ascent are observed to the west and easterly winds and descent to the east of the warm phase of the Kelvin wave, which tilts upward above ~200 hPa from ~60°E to ~150°E (Fig. 1.6a), above the region of suppressed convection (Fig. 1.2d). The associated geopotential height signature is approximately in phase with $u$ (Fig. 1.6b). This configuration of temperature and circulation anomalies is consistent with the idealized Kelvin-wave pattern described in Holton and Lindzen (1968), Fig. 7 of Wallace and Kousky (1968), and Holton (1979).

MJO-related anomalies in clouds and water vapor concentrations in the TTL are shown in Fig. 1.7 and 1.8. Observations of TTL cirrus from the CALIPSO satellite, represented in Fig. 1.7a as anomalies of cloud fraction with bases above 15 km, exhibit close agreement with the 100-hPa temperature anomalies. Suppressed convection is indicated by the 300-hPa cloud minimum near 100°E in Fig. 1.8a, and a region of suppressed cirrus extends eastward and upward from it, coincident with the band of anomalously high temperatures. Enhanced cirrus is
observed over the equator in the central Pacific and extends westward into the subtropics in association with the Rossby wave features (Fig. 1.7a), as indicated by analyses of cirrus clouds based on data from Cryogenic Limb Array Etalon Spectrometer (CLAES; Eguchi and Shiotani 2004) and CALIPSO (Virts and Wallace 2010). The fast-propagating Kelvin wave front circumnavigating the equator gives rise to TTL cirrus anomalies above equatorial South America and Africa. No cirrus signal is evident over central America or subtropical South America, where the climatological-mean relative humidity in the TTL is too low to support widespread cirrus (Virts et al. 2010).

The abundance of water vapor in the atmosphere can be represented by the nearly equivalent entities specific humidity, which is available in the ERA-Interim output, and water vapor mixing ratio, which is measured by MLS. The RMM2 anomaly patterns of water vapor mixing ratio and specific humidity at 100-hPa, shown in Fig. 1.7b, c, are in close agreement, even though MLS water vapor data are not input to the ERA-Interim reanalyses (Dee et al. 2011). Low water vapor concentrations are observed over the central Pacific and extend westward at ~10° latitude toward northern Australia and the Philippines, in agreement with previous analysis of MLS data by Eguchi and Shiotani (2004). Both condensation of water vapor to form cirrus clouds and the strong temperature dependence of saturation water vapor mixing ratio at the extremely low temperatures observed in the TTL contribute to the anomalously low water vapor concentrations in these areas. The temperature, cloud, and water vapor mixing ratio anomaly fields are not precisely aligned, however, as indicated by the displacement between the center of the warm anomaly at ~110°E above the equator and the water vapor mixing ratio anomaly approximately 20° to 30° of longitude to the west of it. The relationship between temperature and water vapor in the TTL is more clearly seen in the cross-
section of RMM2 MLS mixing ratio anomalies in the equatorial belt in Fig. 1.8b. Above the level of main convective outflow (~250 hPa; Highwood and Hoskins 1998), the temperature and water vapor anomalies above the Indian Ocean and Maritime Continent are nearly in quadrature—the cold phase of the Kelvin wave that is propagating downward and eastward is followed ¼ cycle later by a region with anomalously low water vapor concentrations, and vice versa. As air layers ascend and undergo adiabatic cooling in the cold phase of the Kelvin wave, water vapor condenses, forming cirrus clouds. After the layer reaches its lowest temperature, the dehydrated air can increase in water vapor concentration through the evaporation of the cirrus or through advection. Hence, the water vapor anomalies lag the temperature anomalies by ~1-2 weeks, in agreement with results of Schwartz et al. (2008), based on MLS data.

Animations of 100-hPa water vapor anomalies during the MJO cycle, shown in the on-line supplementary material, illustrate the propagation of the water vapor signal eastward over equatorial South America and Africa; however, the anomalies in these areas are nearly a factor of two smaller in magnitude than those above the equatorial eastern Indian Ocean and the western Pacific.

Climatological-mean relative humidity in the TTL is high but not saturated, as shown in the reanalysis fields in Virts and Wallace (2010). Because relative humidity is the ratio between the actual and saturation water vapor mixing ratio, the MJO temperature and water vapor perturbations described above produce opposite tendencies in the relative humidity. For example, above the central Pacific at the time of maximum RMM2, the anomalously low temperatures would tend to raise the relative humidity, while anomalously low water vapor concentrations would tend to lower the relative humidity. The 100-hPa relative humidity
anomaly pattern shown in Fig. 1.7d closely resembles the temperature pattern throughout the tropical belt, with sign reversed, indicating the dominant influence of the temperature anomalies.

Climatological-mean ozone mixing ratios increase rapidly with height from the tropical upper troposphere to the lower stratosphere (Takashima and Shiotani 2007), where ozone is formed through the photolysis of O₂ molecules. Thus, ascent within the TTL and lower stratosphere brings up air with lower ozone concentrations (Randel 1990; Fujiwara et al. 1998; Mote and Dunkerton 2004). The map of filtered MLS 100-hPa ozone mixing ratios regressed onto RMM2, shown in Fig. 1.9a, indicates that anomalously low ozone concentrations are observed in association with cold anomalies (i.e., air layers with a recent history of ascent; Fig. 1.5c), and vice versa. The ozone anomalies are small in the equatorial belt, particularly above the Maritime Continent, but are larger in the subtropics, where climatological-mean ozone concentrations are higher. The Rossby-wave signature in subtropical ozone was also noted in MJO composites of ozone column observations from several satellites (Tian et al. 2007). In contrast, carbon monoxide is produced near the Earth’s surface by processes including biomass burning and fossil fuel combustion (Holloway et al. 2000), is well mixed in the troposphere, and decreases with height in the TTL and lower stratosphere (Fueglistaler et al. 2009). Anomalously high carbon monoxide mixing ratios at the time of maximum RMM2, shown in Fig. 1.9b, are observed in association with low temperatures, and vice versa. Anomalies of both ozone and carbon monoxide are planetary in scale and exhibit the distinctive spatial patterns of Kelvin and Rossby waves, as discussed in Section 1.3a.

Thus, when examining the MJO-related perturbations in Fig. 1.3 to 1.9, a consistent picture emerges in which convection gives rise to planetary-scale regions with air masses with contrasting properties that identify it as “tropospheric”—low temperatures, widespread
cloudiness, and anomalously high carbon monoxide and low ozone concentrations—or “stratospheric” (the reverse).

**Section 1.4. MJO signature in zonal-mean cross-sections**

In the previous section, we investigated the covariability of TTL temperature, circulation, clouds, and trace gases during the MJO, examining the planetary-wave anomalies in each field at the 100-hPa level and in longitude-height cross-sections above the equator. The planetary-wave response to the MJO has been the subject of numerous studies, beginning with Madden and Julian (1972) and continuing in many of the studies cited in Section 1.3. Zonal-mean circulation anomalies associated with the MJO have received comparatively less attention.

In order to analyze the zonal-mean impact of the MJO, filtered tropical-mean (equatorward of 10° latitude) time series of each reanalysis and satellite variable examined in Section 1.3 have been correlated with a series of linear combinations of RMM1 and RMM2. The linear combination with which the filtered time series has the strongest positive correlation (or the strongest negative correlation in the case of temperature and water vapor and ozone concentrations) is shown by the colored lines in the MJO phase diagram in Fig. 1.1. The length of each line indicates the magnitude of the correlation, and variables whose tropical-mean correlations with the MJO exceed the 95% significance level are indicated by solid lines.

It can be seen in Fig. 1.1 that the strongest westerlies, highest heating rates, lowest temperatures, highest relative humidities, and cloudiest conditions at 100 hPa are observed when the MJO convection is moving from the Maritime Continent into the western Pacific. Comparing composites of temperature and TTL cirrus fraction at this stage of the MJO to those observed when the convection is over the eastern Indian Ocean, Virts and Wallace (2010)
reported a decrease in tropical-mean 100-hPa temperature exceeding 2°C and a more than doubling of tropical-mean TTL cirrus fraction. Correlations between tropical-mean trace gas concentrations and the MJO, indicated by the dashed lines in Fig. 1.1, are not statistically significant. It is worth noting, however, that the minimum in tropical-mean ozone is observed at a similar point in the MJO cycle as the minimum in 100-hPa temperature and the maximum in TTL cirrus fractional coverage. Both the observed and reanalysis fields of tropical-mean water vapor concentration indicate a minimum approximately ¼ cycle after the minimum in temperature, or ~10-15 days after dehydration is observed at 100 hPa above the western Pacific (Wong and Dessler 2007, based on MLS data). Tropical-mean carbon monoxide fluctuations exhibit quite different timing and are related to an injection of carbon monoxide-rich air over Africa (Wong and Dessler 2007).

Zonal-mean cross-sections of correlations between RMM2 and filtered COSMIC temperature, ERA-Interim zonal wind and mean meridional circulation, CALIPSO cloud fraction, and MLS water vapor mixing ratio are shown in Fig. 1.10. As indicated by Fig. 1.1, RMM2 represents the stage of the MJO cycle shortly after most of these variables exhibit their largest MJO-related tropical-mean amplitudes. Animations illustrating the evolution of the zonal-mean fields during a typical MJO cycle are available in the on-line supplementary material.

At the time of maximum RMM2, MJO-related convection is enhanced over the western Pacific and suppressed over the eastern Indian Ocean (Fig. 1.2). In the TTL, westerly winds extend from the region of divergence above the central Pacific Ocean eastward for over 240° of longitude, associated with both the Kelvin wave front and the region of convergence above the Maritime Continent (Fig. 1.5a, 1.6). The zonal-mean zonal wind field at the time of maximum
RMM2, shown in Fig. 1.10b, is accordingly dominated by westerly winds that extend from the upper troposphere up to ~100 hPa. Overlying the westerly winds in the TTL is a region of zonal-mean easterly winds in the lower stratosphere.

The mean meridional circulation at the time of maximum RMM2, shown in Fig. 1.10c, exhibits contrasting characteristics below and above the 100-hPa level. Below 100 hPa, the meridional winds are equatorward, converging just south of the equator, near the latitude of the strongest westerly wind anomalies. The radiative heating rates indicate subsidence between ~200 and 100 hPa from the equator poleward out to at least 30° latitude, and the associated adiabatic warming is evident in the observations of anomalously high temperatures in that layer (Fig. 1.10a). Warming of the upper troposphere in association with the fast-propagating Kelvin wave front was previously observed by Bantzer and Wallace (1996), using Microwave Sounding Unit (MSU) channel 3 and 4 temperatures. Convection in the equatorial belt is suppressed at the time of maximum RMM2, as illustrated by the cloudiness anomalies in Fig. 1.10d which are centered just south of the equator, while enhanced cloudiness and associated weak cold anomalies are observed in the lower TTL over the subtropics. The radiative heating rates in the lower portion of the cross-section should be interpreted with caution, because latent heating associated with the convection is expected to have a greater impact there. Above 100 hPa, the meridional winds are also equatorward, and heating rate anomalies are positive up to ~70 hPa and poleward out to ~20° latitude (Fig. 1.10c). The heating rate and temperature anomalies suggest that the strongest zonal-mean ascent tends to be observed around 90-80 hPa, and this is corroborated by the zonally-symmetric TTL cirrus anomalies in Fig. 1.10d that also extend poleward out to ~20° latitude.
The zonal-mean MLS water vapor anomalies in the lower TTL at the time of maximum RMM2, shown in Fig. 1.10e, are consistent with the CALIPSO cloudiness anomalies in Fig. 1.10d—enhanced convection corresponds to anomalously high water vapor concentrations, and vice versa. The low temperatures between 100 and 70 hPa at the time of maximum RMM2 are associated with anomalously low water vapor mixing ratios. In the zonal-mean water vapor animation, which is included in the supplementary material, the water vapor anomalies over the equator appear to shift upward during the evolution of the MJO. This behavior is of questionable significance given that the effective vertical resolution of MLS water vapor profiles near 100 hPa is ~3 km (Read et al. 2007). Similar behavior is not observed in animations of zonal-mean ERA-Interim specific humidity (not shown).

Section 1.5. Conclusions

In this study, we have examined MJO-related variations in TTL temperatures, circulation, clouds, and trace gases. Our results indicate that these fields vary in a physically coherent way during the evolution of the MJO. Perturbations in the TTL are not localized near the anomalous MJO-related convection; rather, each atmospheric field shown in Fig. 1.3 to 1.9 exhibits planetary-scale perturbations that are consistent with the circulation anomalies associated with equatorial Kelvin and Rossby waves. Positive heating rate anomalies coincide with regions with a history of ascent, consistent with radiative relaxation and anomalously low temperatures in a thermally indirect circulation. Adiabatic cooling associated with the wave-driven ascent also gives rise to planetary-scale regions of enhanced relative humidity and TTL cirrus, and vice versa. Water vapor mixing ratios decrease where TTL cirrus are enhanced and lag the cold anomalies by ~1-2 weeks. MJO-related anomalies of the trace gases ozone and carbon
monoxide are consistent with a vertical velocity field acting on a strong climatological-mean vertical gradient.

The zonal-mean signature of the MJO at the time of maximum RMM2 (Fig. 1.10) consists of westerly wind anomalies throughout much of the TTL and easterly anomalies in the lower stratosphere, accompanied by meridional convergence into the equatorial belt. The 100-hPa level marks the approximate transition level between subsidence, adiabatic warming, and suppressed convection below and ascent, cold anomalies, and enhanced TTL cirrus coverage above. The polarity of the zonal-mean anomalies above the ~150-hPa level appears to be determined by the polarity of the planetary-wave perturbations over the central Pacific, the Americas, and Africa, and is opposite to the perturbations over the Maritime Continent, as can be verified by comparing Fig. 1.10 with Fig. 1.3 to 1.8. Put another way, the zonal-mean RMM2 signature is dominated by the fast-moving Kelvin wave front that circumnavigates the equatorial belt.

The MJO is associated with a distinctive set of planetary-scale anomalies in the TTL, many of which have been analyzed in previous studies that were cited above. What is new is the use of the MJO as a carrier signal—that is, we have analyzed MJO variations based on reanalysis fields as well as a suite of independent, satellite-based observations in order to gain a cohesive view of the large-scale relationships among temperature, circulation, clouds, and trace gas concentrations in the TTL. Gettelman et al. (2001) analyzed ENSO variations in a corresponding fashion. The MJO is well-suited to this type of analysis because it dominates tropical variability on its characteristic timescale and has experienced enough realizations during the relatively short satellite record to permit statistical analysis.
In addition to giving insight on the physical relationships among these atmospheric fields, as summarized above, this analysis also offers a comparison between reanalysis fields, which assimilate observations but are also model-dependent, and independent observations. COSMIC temperature profiles are assimilated into the ERA-Interim, as are MLS ozone profiles, but CALIPSO and MLS water vapor and carbon monoxide mixing ratios are not (Dee et al. 2011). The close correspondence between anomalies of CALIPSO TTL cirrus fraction (Fig. 1.7a) and of ERA-Interim heating rate (Fig. 1.5c) and relative humidity (Fig. 1.7d), and between anomalies of MLS water vapor mixing ratio (Fig. 1.7b) and of ERA-Interim specific humidity (Fig. 1.7c), offers reassurance that ERA-Interim analyses representation of the TTL is valid.

Acknowledgments

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Figure 1.1. Phase space defined by two components of Real-time Multivariate MJO index; modified from Fig. 7 of Wheeler and Hendon (2004), but retaining their indicated regions of enhanced MJO-related convection. Colored lines indicate phase and amplitude of strongest correlation between the MJO index and 80-day high-pass filtered tropical-mean (10°S to 10°N) time series of the indicated variables (here, Q is the radiative heating rate, and CF is the TTL cirrus cloud fraction). Solid (dashed) lines indicate that correlations are (are not) statistically significant at the 95% level.
Figure 1.2. 80-day high-pass filtered OLR regressed onto RMM1-RMM2 (a), RMM1 (b), RMM1+RMM2 (c), and RMM2 (d). Contour interval (CI) = 2 W m\(^{-2}\); 0 W m\(^{-2}\) contour is in red.
Figure 1.3. Filtered COSMIC 100-hPa temperature regressed onto RMM1-RMM2 (a), RMM1 (b), RMM1+RMM2 (c), and RMM2 (d). CI = 0.1 K; 0 K contour is in red.
Figure 1.4. Filtered COSMIC temperature (averaged from 5°S to 5°N) correlated with RMM1-RMM2 (a), RMM1 (b), RMM1+RMM2 (c), and RMM2 (d). CI = 0.1; 0 contour is in red.
Figure 1.5. Filtered COSMIC 100-hPa temperature (contours; CI = 0.1 K; 0 K contour is in red), ERA-Interim 100-hPa zonal and meridional wind (a, vectors), vertical velocity in pressure coordinates (b, colors, Pa s$^{-1}$), and ERA-Interim heating rate (c, colors, K day$^{-1}$) regressed onto RMM2.
Figure 1.6. Filtered ERA-Interim zonal wind and heating rate (vectors in both panels), COSMIC temperature (contours in $a$; CI = 0.1; 0 contour is in red), and ERA-Interim geopotential height (contours and colors in $b$; CI = 0.1; 0 contour is in red) correlated with RMM2. Variables are averaged from 5ºS to 5ºN.
Figure 1.7. Filtered COSMIC 100-hPa temperature (contours; CI = 0.1 K; 0 K contour is in red), CALIPSO TTL cirrus index (a, colors), MLS 100-hPa water vapor mixing ratio (b, colors), and ERA-Interim 100-hPa specific humidity (c, colors) and relative humidity (d, colors), and regressed onto RMM2.
Figure 1.8. Filtered COSMIC temperature (contours; CI = 0.1; 0 contour is in red), CALIPSO height-dependent cloud index \((a\), colors\), and MLS water vapor mixing ratio \((b\), colors\) correlated with RMM2. Variables are averaged from 5°S to 5°N.
Figure 1.9. Filtered COSMIC 100-hPa temperature (contours; CI = 0.1 K; 0 K contour is in red) and MLS 100-hPa ozone mixing ratio ($a$, colors) and carbon monoxide mixing ratio ($b$, colors) regressed onto RMM2.
Figure 1.10. Filtered zonal-mean COSMIC temperature (contours in all panels and colors in a; CI = 0.1; 0 contour is in red), ERA-Interim zonal wind (b, colors), meridional wind, and heating rate (c, vectors), CALIPSO height-dependent cloud index (d, colors), and MLS water vapor mixing ratio (e, colors) correlated with RMM2.
Chapter Two: Daily and intraseasonal relationships between lightning and NO$_2$ over the Maritime Continent


Abstract

The relationship between lightning and NO$_2$ over Indonesia is examined on daily and intraseasonal time scales based on lightning observations from the World Wide Lightning Location Network (WWLLN) and tropospheric NO$_2$ column densities from the Global Ozone Monitoring Experiment (GOME-2) satellite mission. Composites of the daily NO$_2$ observations regressed onto lightning frequency reveal a plume of enhanced NO$_2$ following a day of enhanced lightning. Lightning and NO$_2$ also vary coherently with the intraseasonal Madden-Julian Oscillation (MJO) in a manner distinct from the cloudiness signature, with variations of up to ~50% of the annual mean.

Section 2.1. Introduction

Nitrogen oxide radicals (NO$_x$ = NO + NO$_2$) catalyze ozone production in the troposphere, particularly in regions removed from strong local pollution [Davis et al., 1996], regulate the partitioning of hydrogen oxide radicals [Logan et al., 1981; Shindell et al., 2009], and, in the case of NO$_2$, absorb visible radiation, producing local radiative heating [Solomon et al., 1999]. Fossil fuel combustion, biomass burning, and soil emissions are major sources of tropospheric NO$_x$ [Schumann and Huntrieser, 2007]. Lightning is also an important natural source of NO$_x$, 33
estimated to be on the order of 1-10 Tg N yr$^{-1}$ globally [Schumann and Huntrieser, 2007; see also Boersma et al., 2005; Lamarque et al., 1996; Martin et al., 2007; Bucsela et al., 2010], or about 5-10% of the present-day global tropospheric NO$_x$ source [Jaegle et al., 2005]. However, a recent study by Beirle et al. [2010] that directly compared satellite retrievals of NO$_2$ with areas of high lightning activity observed production efficiencies a factor of two or more lower than the lowest estimates from previous studies.

The lightning NO$_x$ source remains uncertain for many reasons, including the physics and detection of lightning itself and the difficulty in detecting lightning NO$_x$ and distinguishing it from other NO$_x$ sources, such as surface combustion [Beirle et al., 2010]. Here we provide a unique observational analysis of the relationship between tropospheric NO$_2$ vertical column density and lightning frequency near Indonesia based on regression analysis performed on daily data and composite maps for various phases of the Madden-Julian Oscillation (MJO).

Section 2.2. Data

Based on an analysis of data from the SCIAMACHY satellite by Beirle et al. [2010], we expect the correlation between the frequency of lightning strokes and NO$_2$ to be weak, requiring a large sample size to detect a statistically significant signal. Accordingly, we chose to utilize data from the Global Ozone Monitoring Experiment (GOME-2) with its larger areal coverage, which enables us to sample the NO$_2$ response to 4-5 times as many lightning strokes as would be possible using SCIAMACHY data (see Section A2.1 in the appendix to chapter two for a discussion of the statistical significance of our results).

The GOME-2 scanning spectrometer provides slant-path, total NO$_2$ retrievals, from which tropospheric NO$_2$ vertical column densities, hereafter referred to as NO$_2$ columns or
simply as NO$_2$, are extracted [Boersma et al., 2004]. The Fast Retrieval Scheme for Clouds from the Oxygen A band (FRESCO) algorithm derives radiance-based effective cloud fractions from GOME-2 retrievals, and a different algorithm indicates whether the tropospheric NO$_2$ retrieval was meaningful [Koelemeijer et al., 2001]. The collocation of FRESCO cloud fraction with GOME-2 NO$_2$ observations allows testing for the effects of cloud contamination.

The World Wide Lightning Location Network (WWLLN) is a network of 68 detector stations that monitor very low frequency radio waves called lightning sferics. The time of group arrival of the radiated waveform is used to locate lightning to within ~5 km. The average detection frequency for all lightning strokes over Indonesia during the period of overlap with GOME-2 can be represented to first order as ~10% [Rodger et al., 2009; see also Section A2.2]. Of the detected WWLLN strokes the median stroke power in the area of this study was $2 \cdot 10^6$ W at the end of 2009 [Hutchins et al., 2010], which is low enough that nearly all lightning-producing storms are detected [Jacobson et al., 2006].

The NO$_2$ and lightning datasets were gridded and averaged as follows. GOME-2 tropospheric NO$_2$ retrievals flagged as not meaningful by the GOME-2 quality control algorithms were discarded, and the remaining data from April 2007 to October 2010 were averaged onto a daily 1° latitude × 1° longitude grid; similar averaging was performed for FRESCO cloud fractions. GOME-2 overpasses are at 9:30 LT; i.e., at 0130 UTC over the western part of our domain and 2330 UTC of the previous day in the eastern part of our domain (10°S to 10°N; 90°E to 150°E). WWLLN strokes were counted daily (from 0200 UTC to 0159 UTC so that each “day” ends at approximately the time of the GOME-2 overpass) in each 1° × 1° bin of the study area for January 2005 to October 2010. An 80-day high-pass Lanczos filter
was applied to the daily time series of each variable, as indicated, to remove the poorly sampled low frequency variability.

In Section 2.4, the Madden-Julian Oscillation is represented by the daily Real-Time Multivariate MJO index developed by *Wheeler and Hendon* [2004], which is based on the dominant modes of variability in near-equatorial outgoing longwave radiation (OLR, a measure of the prevalence of clouds with high, cold tops) and zonal winds in the lower (850 hPa) and upper (200 hPa) troposphere.

**Section 2.3. Seasonal-mean and daily variability**

Seasonal-mean cloud fraction, lightning frequency, and tropospheric NO\(_2\) column over the Maritime Continent are shown in Fig. 2.1. Lightning tends to be more frequent in the summer hemisphere, with maxima along the coasts of the larger islands. In contrast, the NO\(_2\) seasonal cycle is anchored to sources such as urban areas and ship tracks (e.g., extending west from the northern tip of Sumatra during DJF [*Richter et al.*, 2004]).

Coupled day-to-day variability in lightning and NO\(_2\) is revealed by lag-regression analysis of daily, 80-day high-pass filtered data. The 100 1° × 1° grid boxes with the highest annual-mean lightning frequency are designated as “reference grid boxes;” most are over land, but similar results are obtained using ocean reference boxes (see Section A2.3). NO\(_2\) time series for each grid point within a 20° × 20° box centered on the reference grid point are regressed onto time series of lightning at the reference grid point summed over the 24 hours before and after the GOME-2 overpasses. The 100 regression maps for NO\(_2\) observations prior and subsequent to the lightning observations are then centered on their respective reference grid boxes and averaged to form the composite lag regression maps shown in Fig. 2.2.
NO$_2$ observations subsequent to days of enhanced lightning reveal enhanced NO$_2$ around the reference point. The westward shift of the NO$_2$ feature relative to the reference grid box likely reflects advection by winds in the middle and upper troposphere, which have an easterly component during all months. The weaker NO$_2$ maximum near the reference grid box on the day prior to the lightning observations is likely due to day-to-day autocorrelation of both lightning frequency and NO$_2$. Similar patterns are obtained when the analysis is performed with NO$_2$ observations for which the FRESCO cloud fraction was < 0.1 (see Section A2.3). Hence, cloud contamination does not appear to be a significant issue.

An accurate determination of the average impact of a single lightning stroke on NO$_x$ in the surrounding region requires a detailed accounting of the WWLLN detection efficiency, systematic uncertainties in the NO$_2$ retrievals, and consideration of NO$_x$ partitioning, chemistry, and transport on the timescale of ~1 day, all of which are beyond the scope of this paper. However, as a check on whether the above results are reasonable, NO$_2$ production in this region can be roughly estimated from the subsequent regression map. Summing the regression coefficients for both lightning and NO$_2$ within 5° relative latitude and longitude of the reference boxes and assuming that WWLLN detected ~10% of the lightning strokes over Indonesia during the period of overlap with GOME-2, we estimate between $1.7$ and $2.5 \times 10^{25}$ NO$_2$ molecules are produced per stroke regardless of whether we use all retrievals classed as meaningful in accordance with the FRESCO criterion or whether we sample only retrievals with cloud fractions < 0.1. Our estimates also fall within this range if we sample only reference grid boxes that lie over water. Further specifics of these sensitivity tests are reported in Section A2.3. If we use results from Beirle et al. [2009] for estimating the production of NO$_x$ per lightning stroke (LNO$_x$) from satellite observations of NO$_2$, our LNO$_x$ estimates generally lie on the lower end of
the range of 0.2 to $4.0 \times 10^{26}$ molecules per stroke summarized previously [Schumann and Huntrieser, 2007, and references therein] and within the range of values reported by Beirle et al. [2010] based on an analysis of the WWLLN data in conjunction with individual overpasses of the SCIAMACHY instrument. All else the same, these lower LNO$_x$ values would correspond to a global LNO$_x$ source $\sim 1.5$ Tg N yr$^{-1}$ for an average global lightning frequency of 44 s$^{-1}$ [Schumann and Huntrieser, 2007; Beirle et al., 2009].

Section 2.4. Intraseasonal variability

The Madden-Julian Oscillation [Zhang, 2005] dominates tropical atmospheric variability at intraseasonal time scales (30-90 day periods). Its dominant features are often represented as a cycle made up of eight phases. A characteristic feature of the MJO, shown in Fig. 2.3, is an area of enhanced precipitation that develops over the Indian Ocean during phase 1 and propagates eastward across Indonesia during phases 3 and 4. By construction, the MJO precipitation pattern during the remaining phases (5-8) is identical to those for phases 1-4 but with sign reversed.

MJO cloud, lightning, and NO$_2$ patterns over Indonesia are shown in Fig. 2.4. The regression coefficient at each grid point has been divided by the annual mean for that point so that, for example, a lightning regression coefficient of 0.5 indicates an MJO perturbation equivalent to 50% of the climatological mean lightning frequency at that point.

In MJO phases 1 and 2, cloudiness is enhanced west of Sumatra and over western Borneo (Fig. 2.4). More extensive clouds are present in phase 3 and cover the marine portion of the domain by phase 4. In association with the enhanced cloudiness, lightning frequency and NO$_2$ concentrations are also enhanced west of Sumatra and, less prominently, over western Borneo during MJO phase 1. However, lightning and NO$_2$ both decrease as MJO-related cloudiness and
precipitation develop over the study area in MJO phases 2-4. A decrease in lightning during the period of enhanced MJO precipitation has previously been noted in analyses of data from the Tropical Rainfall Measuring Mission Lightning Imaging Sensor [Morita et al., 2006; Kodama et al., 2006]. MJO-induced variations in NO$_2$ have not, to our knowledge, been discussed, and the patterns in Fig. 2.4 cannot be explained by MJO-related variations in tropopause height (see Section A2.5). With the exception of western Borneo, the strongest and most coherent MJO cloud, lightning, and NO$_2$ signals in Fig. 2.4 (and in corresponding correlation maps, see Section A2.4) are observed over marine areas and not over the areas that exhibit a particularly high annual-mean frequency of occurrence of lightning in Fig. 2.1. The MJO-related variations in NO$_2$ columns and lightning frequency range up to ±50% of their respective annual mean values.

Section 2.5. Conclusions

Elevated NO$_2$ concentrations observed following days of frequent lightning suggest that lightning is an important source of NO$_x$ over Indonesia. The similar evolution of the lightning and NO$_2$ fields during the MJO offers further confirmation of their covariability. Indeed, it is difficult to imagine how day-to-day changes in surface NO$_2$ emissions and transport could conspire to create the large scale pattern of NO$_2$ variability observed in association with the MJO, with perturbations ranging up to 50% of the climatological mean. Such large intraseasonal variability in tropospheric NO$_x$ has implications for oxidants and other trace gases, which could be explored and quantified in conjunction with the methodology used in this paper.

Acknowledgments
Tropospheric NO\textsubscript{2} column data from the GOME-2 sensor were obtained from www.temis.nl. Lightning location data were provided by WWLLN (http://wwlln.net), a collaboration among over 50 universities and institutions.
Figure 2.1. Mean cloud fraction (left), lightning frequency (middle, strokes day\(^{-1}\) deg\(^{-2}\)), and tropospheric NO\(_2\) column (right, 10\(^{15}\) molecules cm\(^{-2}\)) during December-February (DJF, top) and June-August (JJA, bottom). Lightning color bar is limited to lower values. In upper left panel, S and J indicate locations of Singapore and Jakarta, respectively.

Figure 2.2. Composites of daily NO\(_2\) field regressed onto daily lightning time series at the 100 grid points with highest lightning frequency (units are molecules cm\(^{-2}\) stroke\(^{-1}\); both variables have been 80-day high-pass filtered); regressions are performed for NO\(_2\) overpasses prior to and following lightning observations. See text for details.
Figure 2.3. 80-day high-pass filtered 5° latitude × 5° longitude Global Precipitation Climatology Project precipitation (mm day⁻¹) regressed onto time series representing phases 1-4 (top to bottom) of MJO index. Adapted from Virts and Wallace [2010].
Figure 2.4. 80-day high-pass filtered cloud fraction (left), lightning (middle), and NO\textsubscript{2} (right) regressed onto time series representing MJO phases 1 (top) to 4 (bottom). Regression coefficients are scaled by the annual mean; gray shading indicates areas of low annual mean tropospheric NO\textsubscript{2} column density, where the MJO NO\textsubscript{2} signal is noisy and not statistically significant.
Appendix to Chapter Two

Section A2.1. Statistical significance of correlations

We elected to analyze GOME-2 NO$_2$ retrievals rather than SCIAMACHY retrievals in hopes that the larger sample size provided by GOME-2’s larger footprint would more than compensate for the larger r.m.s. error of the GOME-2 retrievals. Using 3.5 years of daily GOME-2 observations in 1281 $1^\circ \times 1^\circ$ grid boxes extending over the domain of the analysis provides over $10^6$ observations. The effective number of temporal degrees of freedom of the NO$_2$ observations for a given grid point, calculated using the formula of Leith [1973], is $\geq 650$. In order to determine the statistical significance of composite correlation coefficients similar to the composite regression coefficients in Fig. 2.2, it is also necessary to consider the spatial dependence of the reference grid boxes. Using the eigenvalue method described by Bretherton et al. [1999], the effective number of spatial degrees of freedom of the 100 reference grid boxes used to construct Fig. 2.2 is $\geq 50$. Thus, the total effective number of degrees of freedom for composite correlation coefficients between daily NO$_2$ concentrations and lightning flashes in the $1^\circ \times 1^\circ$ grid boxes that were used to construct the regression maps in Fig. 2.2 is $\geq 30000$. Based on the Student’s $t$ test, and assuming a two-sided distribution, this implies that composite correlation coefficients of 0.02 are significant at the 99% level. The observed composite correlations are on the order of 0.05 to 0.07. The high level of significance of the correlations is evidenced by the smoothness of the patterns.

Section A2.2. Further specifics on the performance of WWLLN
Daily WWLLN lightning observations summed over the Indonesian domain (10°S to 10°N; 90°E to 150°E) for the period of overlap with GOME-2 are shown in Fig. A2.1. The number of detected lightning strokes has increased as more stations have been added to the network (e.g., the average detection efficiency for all lightning strokes in the United States has increased from around 3% in 2005 to 11% in 2010 [Jacobson et al., 2006; Abarca et al., 2010]).

Rodger et al. [2009] compared lightning stroke observations from WWLLN and the New Zealand Lightning Detection Network (NZLDN) from 2007 and found that WWLLN detected 10% or more of all strokes over about 25 kA. Given the reasonably close proximity of New Zealand to our study area, it is reasonable to take 10% as a first-order estimate of WWLLN’s detection efficiency over Indonesia during the period of overlap with GOME-2. Patterns very similar to those shown in Figs. 2.1, 2.4, and A2.5 are obtained when the analysis is performed on lightning data restricted to the period of overlap with GOME-2.

WWLLN’s stroke detection efficiency has been compared in detail to simultaneous observations by the far more sensitive Los Alamos Sferic Array [Smith et al., 2002]. The WWLLN detection efficiency was found in this manner to depend primarily on stroke current amplitude, and not directly on stroke type [Jacobson et al., 2006]. Thus, although not a “total lightning” detection system, WWLLN is biased toward high current alone, and has no overt selection for ground strokes over cloud strokes. To that extent, WWLLN might be expected to serve as a reasonable interim proxy for lightning NOx production, at least for the period before total-lightning monitors are operational. WWLLN data do not provide any indication of stroke altitude.

The median global very low frequency (VLF) stroke power measured by WWLLN is $3 \times 10^6$ W, more than three orders of magnitude smaller than that indicated by previous
measurements, which have shown the power radiated by strokes to be near $10^{10}$ W [Krider and Guo, 1983]. However, this apparent discrepancy is due to the difference in methodology. Previous measurements were of broad band peak power taken in the near field (within 100 km of the lightning stroke), whereas WWLLN measures the r.m.s. power in the 6-18 kHz band in the far field. With these factors accounted for, the median power is comparable to the previously reported value of $10^{10}$ W [Krider and Guo, 1983].

Section A2.3. Sensitivity tests on daily lightning/NO$_2$ relationship and NO$_2$/stroke estimate

Figure 2.2 shows the results of a composite lag regression analysis of daily lightning frequency and tropospheric NO$_2$ column density fields over Indonesia (see text for details). The composite analysis was based on the reference boxes indicated in Fig. A2.2. To test the robustness of these results, we have conducted a series of sensitivity studies, varying the following aspects of the analysis (note that for each estimate of NO$_2$ production, a WWLLN detection efficiency of 10% was assumed).

- **Cloud fraction threshold** The results shown in Virts et al. [2011] were obtained by analyzing only tropospheric NO$_2$ retrievals classified as “meaningful” by the GOME-2 quality control algorithms [Boersma et al., 2004]. To test the impact of cloudy observations, the composite lag regression analyses were repeated with NO$_2$ time series that incorporated only observations for which the FRESCO cloud fraction [Koelmeijer et al., 2011] was below 0.1 (Fig. A2.3, top row).

- **Sample size** Although it is clearly desirable to choose reference boxes for which there is sufficient lightning to provide a day-to-day signal, our selection of the 100 grid boxes with the highest annual-mean lightning frequencies is somewhat arbitrary. Lag
regression analyses performed on reference boxes with lightning frequencies in the top 50 and 500 are shown in the second and third rows of Fig. A2.3.

- **Location of reference boxes** The reference boxes used in the analysis in Fig. 2.2 are located over land or along the coasts of the Indonesian islands (Fig. A2.2). To test whether our results are sensitive to possible changes in the NO₂ plume from surface sources along the islands, the lag regression analysis was repeated using a set of reference boxes located over water or over less polluted land areas, as indicated in Fig. A2.4. The associated composite NO₂ regression patterns are shown in the bottom row of Fig. A2.3.

Comparison of Fig. A2.3 and Fig. 2.2 in the text shows that the spatial pattern and temporal evolution of the NO₂ field are robust with respect to variations in the analysis protocol. The production efficiencies estimated on the basis of these various protocols were used to obtain the estimated range of 1.7 to 2.5 × 10²⁵ NO₂ molecules per stroke put forth in *Virts et al.* [2011].

**Section A2.4. MJO correlation maps**

Figure A2.5 shows maps of correlation coefficients between the MJO index and clouds, lightning, and NO₂, analogous to the maps of regression coefficients shown in Fig. 2.4 in the text. It is evident that the MJO signal in lightning and NO₂ over the eastern Indian Ocean is strongest just to the west of Sumatra, within the region in which mesoscale circulations driven by land/sea contrasts and terrain play an important role in triggering convection. In the correlation maps a weak enhancement of lightning is also apparent over New Guinea in phases 1 and 2, which is mirrored in the NO₂ regression pattern.

**Section A2.5. Alternative explanations for NO₂ patterns**
In this section we examine three factors that can influence tropospheric NO$_2$ column retrievals and discuss whether they could produce NO$_2$ patterns similar to those shown in *Virts et al.* [2011].

- **Tropopause height variations.** Tropospheric NO$_2$ vertical column densities are obtained from total atmospheric slant column densities by subtracting an assumed stratospheric NO$_2$ column and dividing by a tropospheric air mass factor [Boersma et al., 2004]. The MJO modulates cold-point tropopause height; however, these variations are on the planetary scale [e.g., Zhang, 2005; Virts and Wallace, 2010] and thus cannot explain the more localized NO$_2$ patterns in Figs. 2.4 and A2.5.

- **Transport of non-lightning NO$_2$.** NO$_2$ produced by surface sources can be transported by convection to the upper troposphere, where its lifetime is longer and where it is more readily visible to the satellite. Winds can also transport NO$_2$ horizontally, as seen in Fig. 2.2 in the text. We have demonstrated that the plume of enhanced NO$_2$ associated with a lightning maximum is present regardless of whether we use reference boxes over the ocean or over less polluted land areas (Fig. A2.4). In addition, the MJO NO$_2$ signature in Figs. 2.4 and A2.5 is strongest to the west of Sumatra, where the influence of surface sources of NO$_2$ is much lower (e.g., Fig. 2.1).

- **Cloud contamination.** The dissimilarities in the cloud and NO$_2$ patterns in Figs. 2.1 and 2.4, combined with the fact that the NO$_2$ patterns in Figs. 2.2 and 2.4 do not change when only NO$_2$ retrievals with cloud fractions below 0.1 are included in the analysis (see, e.g., Fig. A2.3), indicate that cloud contamination is not a significant issue for these analyses, though it may need to be taken into account in quantifying the lightning NO$_x$ source.
Thus, while each of these factors influences GOME-2’s tropospheric NO$_2$ retrievals, none can account for the NO$_2$ patterns associated with lightning shown in the text. Other factors that impact tropospheric NO$_2$ retrievals are discussed in Boersma et al. [2004].
Figure A2.1. Time series of daily lightning strikes detected by WWLLN over Indonesia.

Figure A2.2. Map of 100 grid boxes with highest annual-mean lightning frequency. These reference boxes were used to generate the composites in Fig. 2.2.
Figure A2.3. As in Fig. 2.2, but regressions were calculated using NO$_2$ time series based on only those observations with FRESCO cloud fractions less than 0.1 (top row), the indicated number of reference boxes (second and third rows), and the reference boxes shown in Fig. A2.4 (bottom row).
Figure A2.4. Annual-mean lightning frequency in selected grid boxes over water or less polluted land areas.

Figure A2.5. As in Fig. 2.4, but cloud, lightning, and NO$_2$ are correlated with MJO indices.
Chapter Three: Highlights of a new ground-based, hourly global lightning climatology

(Based on the following publication: Virts, K. S., J. M. Wallace, M. L. Hutchins, and R. H. Holzworth, Bulletin of the American Meteorological Society, submitted.)

Abstract

The seasonally and diurnally-varying frequency of lightning flashes provides a measure of the frequency of occurrence of intense convection and, as such, is useful in describing the Earth’s climate. Here we present a few highlights of a global lightning climatology based on data from the ground-based World Wide Lightning Location Network (WWLLN), for which global observations began in 2004. Because WWLLN monitors global lightning continuously, it samples ~100 times as many lightning strokes/flashes per year as the Tropical Rainfall Measuring Mission’s (TRMM) Lightning Imaging Sensor (LIS). Using WWLLN data it is possible to generate a global lightning climatology that captures seasonal variations, including those associated with the mid-latitude storm tracks, and resolves the diurnal cycle, thereby illuminating the interplay between sea breezes, mountain-valley wind systems, and remotely forced gravity waves in touching off thunderstorms in a wide variety of geographical settings. The text of the paper shows a few samples of WWLLN-based regional seasonal (the mid-latitude storm tracks and the Mediterranean) and diurnal climatologies (the Maritime Continent, the central Andes, and equatorial Africa), and the on-line supplement presents animations of the global seasonal cycle and of the diurnal cycle for the latter regions.

Section 3.1. Introduction
Approximately half of tropical rainfall occurs in heavy, short-lived showers associated with convective updrafts (Lopez 1978; Rickenbach and Rutledge 1998; Johnson et al. 1999), while the other half falls from stratiform clouds, including convective anvils (Schumacher and Houze 2003). Oceanic convection tends to have updraft velocities weaker than those observed in convection over land, where the buoyancy produced by the daytime heating of the boundary layer can lead to updraft velocities of 5 m s\(^{-1}\) or more (Stitch et al. 2002), which is strong enough to produce the charge separation required to produce lightning (Deierling and Petersen 2008). Hence, the frequency of occurrence of lightning serves as a proxy for the frequency of occurrence of vigorous convection. In addition to increasing the buoyancy of boundary layer air, daytime solar heating also leads to local land-sea and mountain-valley wind regimes. These together form the dominant mechanism for producing convection over the tropical landmasses (Kikuchi and Wang 2008).

In contrast with the predominantly boundary-forced thunderstorms in the tropics, thunderstorms and heavy precipitation in the extratropics are known to occur in association with migrating synoptic scale cyclones (Pessi and Businger 2009), but they still occur preferentially over certain favored regions. During winter, cyclones tend to form in the lee of mountain ranges, such as the Andes (Streten and Troup 1973) and Rockies (Zishka and Smith 1980), and over the western oceanic boundary currents and propagate eastward across the oceans, forming so-called “storm tracks” of enhanced cyclone activity (Hoskins and Valdes 1990). Wintertime cyclones are also observed over the Mediterranean Sea (Alpert et al. 1990).

Until recently, lightning climatologies have been based on station data or local lightning networks, most of which are regional or national in scope. Global satellite-based lightning monitoring began in the 1970s (Turman 1978, and references therein; Orville and Spencer 1979),
and statistically significant lightning climatologies became available with the development of the Optical Transient Detector (OTD) and the Lightning Imaging Sensor (LIS; Christian et al. 2000). Datasets derived from these measurements have been used to construct annual-mean and seasonal lightning climatologies (Christian et al. 1999, 2003) and to investigate tropical-mean diurnal lightning variability (Liu and Zipser 2008).

The World Wide Lightning Location Network (WWLLN, see http://wwlln.net) is a ground-based network with global observations beginning in 2004. The WWLLN record is now long enough to support studies of seasonal, diurnal, and synoptic lightning variability over most of the globe. The WWLLN and LIS datasets are described in Section 3.2, and a comparison of the climatological distribution of lightning detected by each sensor is presented in Section 3.3. Seasonal and diurnal variations in lightning frequency observed by WWLLN are discussed in Sections 3.4 and 3.5, respectively, followed by conclusions in Section 3.6.

Section 3.2. Data

The WWLLN network consists of 68 sensors, as of October 2012 (sensor locations are shown in Fig. 3.1), that monitor very low frequency (VLF) radio waves for lightning sferics. The network uses a time of group arrival technique (Dowden et al. 2002) on the detected sferic waveforms to locate lightning to within ~5 km and < 10 µs (Abarca et al. 2010). Comparisons between lightning observations from WWLLN and regional networks indicate that the global detection efficiency of WWLLN is ~10% (Rodger et al. 2006, 2009; Abarca et al. 2010; Connaughton et al. 2010; Hutchins et al. 2012b) of all strokes, which is sufficient to enable WWLLN to detect almost all lightning-producing storms (Jacobson et al. 2006).
Abarca et al. (2010) compared WWLLN hourly lightning frequency over the United States with observations from the National Lightning Detection Network (NLDN) and found some marked differences. The agreement was somewhat better when comparing subsets of each climatology corresponding to strokes with strong peak currents, but the differences were still large enough that they expressed reservations concerning the ability of WWLLN to capture the diurnal cycle. In addition, attenuation of the VLF waves used by WWLLN to locate lightning strokes is strongest during daytime (Hutchins et al. 2012a), introducing a possible bias toward detecting more nighttime than daytime lightning. We will show in Section 3.5 that the WWLLN hourly climatology is generally consistent with prior ground-based studies and with our understanding of the processes that cause lightning to vary systematically with time of day.

OTD was launched with the MicroLab-1 satellite in April 1995 into a 70° inclination orbit (Christian et al. 2003), and LIS is carried on the Tropical Rainfall Measuring Mission (TRMM) satellite, which was launched in 1997 into a 35° inclination orbit (Christian et al. 1999). In this study, we make use of lightning climatologies based on ~13 years of LIS and ~5 years of OTD observations. Annual-mean and hourly-mean climatologies are available at 0.5° and 2.5° spatial resolution, respectively. TRMM also carries a Precipitation Radar (PR) and Visible and Infrared Scanner (VIRS). TRMM rainfall observations are supplemented with data from other satellite-borne microwave imagers and infrared sensors to generate the gridded TRMM 3B42 dataset (Huffman et al. 2007), which is available at 3-hourly temporal resolution and 0.25° spatial resolution.

Given the complexity of lightning, interpreting the differences between lightning climatologies based on observations from different instruments or networks is not straightforward. LIS/OTD and WWLLN rely on fundamentally different detection methods.
WWLLN receivers detect sferics that have propagated in the Earth-ionosphere waveguide and are fully captured within a 1 millisecond window at each station (Dowden et al. 2002). In contrast, LIS and OTD are optical staring imagers that detect momentary changes in cloud brightness caused by lightning; optical transients that are similarly located in space and time are grouped into events referred to as flashes (Christian et al. 2000). Because WWLLN has a relatively high detection threshold for power, it preferentially detects strong cloud to ground strokes and rarely detects and locates multiple strokes within a single flash (Rodger et al. 2004, 2005; Jacobson et al. 2006). Hence, lightning strokes detected by WWLLN should be nearly equivalent to lightning flashes. It follows that comparing climatologies of WWLLN strokes and LIS/OTD flashes should be informative even though, strictly speaking, strokes and flashes are different phenomena. Further specifics on WWLLN may be found in Section A3.1 of the appendix to Chapter 3 and in the peer-reviewed articles listed at http://wwlln.net/publications.

Section 3.3. TRMM LIS versus WWLLN annual-mean lightning climatologies

The frequency of occurrence of lightning, as detected by WWLLN during the years 2008-2011, is compared with LIS/OTD observations in Figs. 3.2a, b. The two climatologies are qualitatively similar, both showing a concentration of lightning over major tropical continents—Africa, southeastern Asia and Australasia, and Central and South America—with strong gradients near the coastlines and features that bear a strong relationship to the underlying topography. For example, lightning is frequently observed in the central United States between the Rocky and Appalachian Mountains. Both lightning climatologies differ substantially from the TRMM rainfall climatology shown in Fig. 3.3a, in which the maxima are located over the oceanic “warm pool” covering the equatorial Indian and western Pacific Oceans and in the
region of the intertropical convergence zone (ITCZ). Lightning also tends to be more geographically focused than rainfall: in the WWLLN and LIS climatologies for the tropical belt (30°N-30°S), half the lightning strokes are observed in 8% of the area, whereas half the rain falls over 22% of the area (not shown). These distinctions illustrate the importance of daytime heating of the atmospheric boundary layer over land in creating the conditions required for the initiation of intense convection, as discussed in the Introduction.

The color bars in Figs. 3.2a, b have been chosen so as to emphasize the similarities between the LIS/OTD and WWLLN lightning climatologies. Differences between the two climatologies are illustrated in Fig. 3.2c, which shows the point-wise ratio of lightning frequency reported by LIS/OTD and WWLLN. For display purposes, this ratio was multiplied by a scaling factor—the global mean WWLLN lightning frequency divided by the global mean LIS/OTD lightning frequency—so that values < 1 indicate proportionally more WWLLN lightning, and vice versa, relative to their respective global means. With a few notable exceptions (e.g., over the Maritime Continent and the southeastern United States), LIS/OTD reports proportionally more lightning over land and less over the oceans than WWLLN. Because WWLLN’s detection efficiency is higher for lightning strokes with higher peak currents (Rodger et al. 2009), the land/sea contrasts in Fig. 3.2c may reflect a tendency for lightning strokes over the oceans to be more powerful than over land. This tendency was previously noted by Rudlosky and Fuelberg (2010) in an analysis of the mean peak currents in negative cloud-to-ground flashes reported by NLDN. The unevenness of the spacing of the WWLLN stations (Fig. 3.1) also contributes to the observed differences between the LIS/OTD and WWLLN climatologies. For example, it is evident from Fig. 3.2c that the LIS lightning climatology places relatively greater emphasis on the maxima in lightning frequency in areas such as Africa and the Himalayas, where WWLLN’s
detection efficiency is lower (Hutchins et al. 2012b). Further analysis of lightning stroke energy, particularly over the eastern United States, is ongoing.

For broad regional comparisons between lightning frequency over different continental regions, such as central America versus India, LIS is clearly the definitive dataset for latitudes up to ~35°. The advantage of WWLLN is that it samples continuously over the whole globe, whereas LIS and OTD sample only when the satellite passes overhead (~0.1% of the time over much of the tropics; Christian et al. 1999, 2003). When detection efficiency (~10% for WWLLN, as discussed in Section 3.2, and ~80-90% for LIS, according to Boccippio et al. 2002) is taken into account, WWLLN samples ~100 times as many strokes per year as LIS. The difference in the number of samples has little effect on the appearance of the annual-mean lightning climatologies shown in Fig. 3.2, but it becomes a more important consideration when considering how finely the data can be disaggregated by synoptic situation and/or by time of day.

In Sections 4 and 5 we describe some regional, WWLLN-based seasonal and diurnal lightning climatologies that reveal, with an unprecedented level of detail, how the occurrence of vigorous convection is maximized in the extratropical wintertime storm tracks and in the tropics where it is shaped by the underlying topography and land-sea distribution. The selected images for prescribed seasons and times of day shown in the text are complemented by 12-month animations of the climatological-mean seasonal cycle and 24-hour animations of the climatological-mean diurnal cycle in the online supplement for this article, and by a more extensive selection of animations on the WWLLN web site.

**Section 3.4. Seasonal dependence**
WWLLN lightning and TRMM precipitation over the northern Pacific and Atlantic Oceans during Northern Hemisphere winter are shown in Fig. 3.4a, b and in Fig. 3.3b, c, respectively; corresponding maps for the southern Atlantic and Indian Oceans during Southern Hemisphere winter are shown in Fig. 3.4c, d and Fig. 3.3d, e. The zonally elongated precipitation bands observed in each ocean basin between ~30° and 45° latitude are associated with the wintertime storm tracks, as described in Section 3.1; the North Atlantic storm track appears to veer poleward at the downstream end. Vigorous convection is also found in the storm tracks, as evidenced by the corresponding lightning bands in Fig. 3.4. The lightning maxima lie along, or a few degrees equatorward of, the axes of maximum rainfall, and suggestions of secondary east-west oriented lightning maxima are evident at subtropical latitudes to the west of Hawaii and to the southeast of Bermuda. Based on data from LIS and the Pacific Lightning Detection Network, Pessi and Businger (2009) concluded that much of the lightning in the storm tracks occurs in thunderstorms embedded in the cold fronts of mid-latitude cyclones. During winter, both lightning densities and rainfall rates are larger over the oceans than over land. Strong gradients are observed along the coasts and over warm western boundary currents. Over Argentina, wintertime cyclone development takes place in the lee of the Andes (Streten and Troup 1973), and the lightning and precipitation maxima both begin over Argentina. Over both the South Atlantic and the Southern Indian Oceans, the lightning tends to be concentrated in the western side of the storm track, where sea surface temperatures are higher. Global, monthly-mean lightning animations based on WWLLN and LIS/OTD observations are in the on-line supplement.

Cyclonic disturbances also produce thunderstorms over the Mediterranean during the local winter season (Defer et al. 2005). Both lightning and precipitation are more frequent over
the warmer sea than over the colder European landmass (Fig. 3.5a, b), in agreement with previous studies based on data from local lightning networks (Altaratz et al. 2003; Katsanos et al. 2007). Lightning gradients along the northern and eastern coasts of the Mediterranean and Adriatic are particularly sharp where there is steep terrain near the coast. Compositing analysis (not shown) shows that southwesterly flow is observed during days of frequent lightning along the northeastern Adriatic coast. The low level instability produced by colder air moving over the warmer Mediterranean waters can also create a favorable environment for thunderstorm development in the absence of cyclones (Altaratz et al. 2003). In contrast, during Northern Hemisphere summer, lightning and precipitation are most frequent over the warmer European continent (Fig. 3.5c, d; Chronis 2012), where thunderstorms form over the Alps, the Pyrenees, and the mountains of the Balkan Peninsula. Lightning occurrence over the latter region was also examined by Kotroni and Lagouvardos (2008) using an experimental lightning network.

Section 3.5. Diurnal dependence

The diurnal cycle of lightning over the Maritime Continent, the central Andes, and the African Great Lakes is summarized in Fig. 3.6-3.8 in terms of maps of the climatological frequency of occurrence of lightning during selected segments of the day. These summary maps and the animations in the supplementary material reveal the following characteristics of the diurnal variability.

• In response to the diurnal cycle in incoming solar radiation, convection begins around local noon over inland regions with relatively flat terrain such as the Amazon basin (Fig. 3.7) and parts of central Africa (Fig. 3.8). Over these regions, lightning occurs most frequently during afternoon and early evening, with a late afternoon peak, and least
frequently during the morning. Rainfall over the tropical continents exhibits a diurnal

cycle with a similarly-timed peak (Kikuchi and Wang 2008). Diurnal lightning

variability is smaller in Argentina (Fig. 3.7), where long-lived mesoscale convective

systems (MCSs; Nesbitt et al. 2000) produce nighttime lightning.

• During sunny days, warm air rises along mountain slopes as an anabatic wind, or “valley

breeze.” Thunderstorms form over the steep terrain just west of the crest of the Andes
during late morning and shift to the east of the crest by ~1600 LT, where they linger into

early evening. The influence of valley breezes is also evident in the intense

thunderstorms that begin in the early afternoon along the western slopes of the Mitumba

and Virunga Mountains of central Africa and produce maximum lightning ~1900 LT

(Fig. 3.8; Jackson et al. 2009). Similarly, afternoon and early evening lightning is

frequently observed over the slopes of the mountains of the Maritime Continent (Fig.

3.6).

• At night, cooling air drains down from the mountains into the valleys as a katabatic wind,
or “mountain breeze,” inducing weak ascent over the lower terrain that is augmented

where valleys of adjacent tributaries converge. Lightning occurs on the valley floor to

the east of the Andes (Fig. 3.7), where nocturnal convection tends to be organized into

mesoscale convective complexes (Bendix et al. 2009), but not on the west side of the

Andes, where the boundary layer air is much drier. Adjacent to the Andes, lightning is

observed most frequently just before midnight and persists into the morning of the next
day. In some cases, nighttime lightning frequencies are similar in magnitude to the
daytime maxima over the mountain slopes. Over the Maritime Continent, the mountains
are free of lightning at night, while thunderstorms linger over the surrounding lowlands (Fig. 3.6).

- Strong diurnal variations in lightning occurrence are also observed near coastlines. On sunny mornings, land surfaces, with their smaller heat capacities, warm up rapidly, resulting in strong temperature and pressure contrasts along coastlines that drive onshore winds, or “sea breezes.” Over the Maritime Continent, convection begins around noon as the sea breeze fronts develop and intensify as they propagate inland during the afternoon, with peak lightning frequencies around 1600-1700 LT (Fig. 3.6). Locally, lightning is enhanced where convex coastlines result in sea breeze convergence; e.g., over parts of Borneo. Most small islands experience brief afternoon lightning maxima ~3-5 hours in duration and become lightning-free by early evening, while thunderstorms over larger islands persist into the evening. Similar behavior has been noted in precipitation duration (Qian 2008). Where mountains and coastlines are in close proximity, sea breezes and valley breezes can act in concert to produce strong afternoon and evening convection (Mahrer and Pielke 1977).

- At night, when the boundary layer over land cools, coastal circulation patterns reverse, resulting in “land breezes.” Thunderstorms begin ~2000 LT over Lakes Malawi and Kivu in central Africa and an hour or two later over Lakes Victoria, Tanganyika, and Albert (Fig. 3.8). Land breezes, enhanced by katabatic winds from the surrounding terrain (Savijärvi and Järvenoja 2000), strengthen and converge over the lakes through the night, and lightning is observed most frequently at ~0400-0500 LT. Over Lake Victoria the lightning maximum shifts from the northeastern part of the lake to the southwestern part by morning; similar behavior was observed by Hirose et al. (2008)
based on TRMM precipitation data. No corresponding shift is observed over the other lakes.

Near the Maritime Continent, the nocturnal convection regime is more complex. Locally, land breezes and mountain breezes result in offshore winds, and areas with concave coastlines exhibit enhanced nighttime lightning (Fig. 3.6). In addition, features in the animations (found in the on-line supplement and at http://wwlln.net/climate) that resemble gravity waves propagate out of the regions of afternoon convection and appear to trigger thunderstorms over the coastal waters, as in the numerical simulations of Mapes et al. (2003). The daytime lightning over the western slopes of the mountains of Sumatra moves offshore in late afternoon into a region of weakening convective inhibition (Wu et al. 2009), propagates southwestward, and weakens after sunrise the following day, in agreement with the description of Mori et al. (2004), based on TRMM rainfall data. Strong convection and frequent lightning occur between 2200 and 1200 LT in the Strait of Malacca, where land breezes from Sumatra and the Malay Peninsula converge (Fujita et al. 2010). Elsewhere, such as along the northern coast of New Guinea, a separate line of thunderstorms forms along the coast just before midnight and moves offshore in the early morning hours. By morning, each of the major islands is surrounded by a ring of lightning, consistent with the TRMM rainfall climatology of Kikuchi and Wang (2008).

Diurnal lightning maps for the Maritime Continent based on the LIS/OTD climatology are shown in Fig. 3.6. The WWLLN and LIS/OTD climatologies are similar in some respects—for example, both indicate that lightning is more frequent over the major islands during the afternoon and evening than during the morning. However, the spatial resolution of the LIS/OTD climatology is too coarse to fully resolve features such as the morning lightning maximum in the
Strait of Malacca. Hour-by-hour time series of WWLLN lightning for three areas of interest are included Section A3.2.

The discussion in this paper has focused on diurnal lightning variability in the climatological-mean or, for the Andes, the extended local summer season. Phenomena such as the monsoons (Kandalaonkar et al. 2003), El Niño-Southern Oscillation (Hamid et al. 2001), and the Madden-Julian Oscillation (Kodama et al. 2006) also modulate lightning occurrence and, likely, its diurnal cycle.

Section 3.6. Conclusions

TRMM LIS and WWLLN are the only existing lightning datasets that provide global coverage. LIS provides a well-calibrated tropical lightning climatology that already extends for over 14 years. WWLLN has been in operation for only a few years, but it offers the possibility of monitoring lightning frequency over the entire globe with sample sizes two orders of magnitude larger than is feasible with LIS, as evidenced by the comparison in Fig. 3.6. The two datasets are highly complementary. The Geostationary Operational Environmental Satellite R-Series (GOES-R) satellite, scheduled for launch in 2015, will carry aboard a lightning sensor that will provide continuous coverage over the North American sector, thereby providing a certain degree of redundancy in the lightning measurements in that sector.

In this paper we have shown some highlights from a seasonal and diurnal cycle lightning climatology constructed from WWLLN data binned hourly and in $0.25^\circ \times 0.25^\circ$ grid boxes. Though no single lightning climatology can be said to be definitive, the WWLLN lightning climatology appears to be consistent with in situ observations and with the TRMM rainfall climatology, and it provides a plausible representation of how the frequency of deep cumulus
convection varies over the course of the day in the vicinity of coastlines and orography. Indeed, most of the features of the seasonal and diurnal cycles derived from WWLLN and the mechanisms that give rise to them have been known, or at least hinted at, for decades. What is new is the unprecedented ability to view the diurnal cycle in lightning from a global perspective with high spatial resolution. Although we have not illustrated it here, the large sample size of WWLLN data makes it possible to refine the analysis in order to investigate how the diurnal cycle in lightning changes with season and in response to day-to-day variations in wind patterns and vertical profiles of temperature and moisture. Such analyses will be useful for refining our understanding of the environmental conditions that give rise to intense convection and for validating numerical models that attempt to simulate the statistics of intense convection in a realistic setting.

Acknowledgments

Lightning location data were provided by WWLLN (http://wwlln.net), a collaboration among over 50 universities and institutions. The LIS/OTD climatologies were obtained from the NASA’s Global Hydrology and Climate Center (http://thunder.msfc.nasa.gov/). TRMM rainfall data were obtained from NASA’s Goddard Earth Sciences Data and Information Services Center (http://mirador.gsfc.nasa.gov/).
Figure 3.1. Location of WWLLN sensors, color-coded according to the date each was established. Stations established prior to 2008 are shown in dark blue; black stars indicate stations established 2012-present.
Figure 3.2. Annual-mean lightning frequency from LIS/OTD (a; flashes km$^{-2}$ yr$^{-1}$) and WWLLN (b; strokes km$^{-2}$ yr$^{-1}$). Ratio of LIS/OTD to WWLLN strokes, scaled by the mean of each dataset (c; gray shading indicates either no LIS/OTD lightning flashes or WWLLN lightning frequency < 0.01 strokes km$^{-2}$ yr$^{-1}$; see text for details). All fields have been averaged on a 1° $\times$ 1° grid. Black contours indicate the 500-m elevation.
Figure 3.3. TRMM annual-mean (a) and seasonal-mean precipitation (b to e; mm day$^{-1}$; $1^\circ \times 1^\circ$ resolution). Black contours indicate the 500-m elevation.
Figure 3.4. WWLLN seasonal-mean lightning frequency (strokes km$^{-2}$ yr$^{-1}$; 1° × 1° resolution) during December-February (DJF; a, b) and June-August (JJA; c, d). Black contours indicate the 500-m elevation.
Figure 3.5. WWLLN lightning frequency (a, c; strokes km⁻² yr⁻¹) and TRMM 3B42 precipitation (b, d; mm day⁻¹; both at 0.25° × 0.25° resolution) during DJF (a, b) and JJA (c, d). Black contours indicate the 500-m elevation.
Figure 3.6. WWLLN lightning frequency (a-c; strokes km$^2$ yr$^{-1}$; hourly, 0.25° × 0.25° resolution) and LIS/OTD lightning frequency (d-f; flashes km$^2$ yr$^{-1}$; two-hourly, 2.5° × 2.5° resolution) for indicated time intervals during all months. Local time given for Singapore. Black contours indicate the 500-m elevation. Elevation (m) in (g).
Figure 3.7. WWLLN lightning frequency (strokes km$^{-2}$ yr$^{-1}$; 0.25° × 0.25° resolution) over the central Andes for indicated time intervals during November-March. Local time given for Lima, Peru. Black contours indicate the 500-m and 4000-m elevation. Elevation (m) in (c).

Figure 3.8. As in Fig. 3.7, but for tropical Africa for all months. Local time given for Kampala, Uganda.
Appendix to Chapter Three

Section A3.1. Further specifics on WWLLN

The number of lightning strokes detected by the World-Wide Lightning Location Network (WWLLN) grows over time as more stations are added to the network, as illustrated in Fig. A3.1, which shows daily WWLLN observations summed over the globe and over the Maritime Continent, the central Andes, and the African Great Lakes, for which the diurnal variability is documented in Figs. 3.6-3.8. Integrated over the globe, the sampling of lightning strokes by WWLLN has roughly doubled over the period of this study. However, in the averages over the regions highlighted in this study, the sampling rate has been dominated by the annual cycle.

Section A3.2. Hourly lightning time series and animations

Hourly, climatological-mean WWLLN lightning frequencies averaged along three reference lines are shown in Fig. A3.2. Over the mountains of Sumatra, along the blue line, thunderstorms tend to initiate around noon, and the maximum lightning frequency is observed in early evening. The mountains are relatively lightning-free at night. In contrast, along the red line, which lies parallel to the coastline, lightning tends to be suppressed during the afternoon, in agreement with subsidence associated with the sinking branch of the sea breeze circulation. The area of enhanced thunderstorms shifts offshore to the eastern Indian Ocean during evening and night, and the associated lightning frequencies can be larger than the daytime maximum over the mountains. The Strait of Malacca, indicated by the green line, exhibits a maximum in lightning frequency during the early morning, as the converging land breezes from Sumatra and the Malay
Peninsula produce strong convergence in the Strait. The diurnal evolution of lightning in these three areas is consistent with previous studies based on in situ observations and on TRMM rainfall observations which are cited in Section 5 of the text.

Elsewhere in the supplementary material are animations of monthly-mean lightning based on the WWLLN and LIS/OTD climatologies, shown at 1° × 1° resolution, and animations of WWLLN hourly lightning for the Maritime Continent, the central Andes, and the African Great Lakes, shown at 0.25° × 0.25° resolution. The WWLLN animations are based on observations from 2008-2011, and all months are included in the diurnal animations. Gray or black contours indicate the 500-m elevation contour, and the 4000-m elevation contour is also shown in the Andes animation. Hourly lightning animations for the globe and for many other regions of the world are available at the WWLLN website (http://wwlln.net/climate/).
Figure A3.1. Time series of daily lightning strokes detected by WWLLN over the indicated regions.
Figure A3.2. Elevation (m) of Maritime Continent (a). WWLLN hourly-mean lightning (strokes km\(^{-2}\) yr\(^{-1}\)) along three reference lines (b). The line colors in (a) match those in (b).
Chapter Four: Diurnal lightning variability over the Maritime Continent:

Impact of low-level winds, cloudiness, and the MJO

(Based on the following publication: Virts, K. S., J. M. Wallace, M. L. Hutchins, and R. H. Holzworth, in preparation)

Abstract

Lightning over the Maritime Continent exhibits a pronounced diurnal cycle. Daytime and evening lightning occurs near coastlines and over mountain slopes, driven by sea and valley breezes. Nocturnal and morning thunderstorms are touched off where land breezes or mountain breezes converge or by gravity waves propagating away from regions of vigorous afternoon convection. In this study, the modulation of the diurnal cycle of lightning and precipitation by 850-hPa winds, cloudiness, and the Madden-Julian Oscillation (MJO) is investigated using observations from the World Wide Lightning Location Network (WWLLN) and the Tropical Rainfall Measuring Mission (TRMM) satellite. 850-hPa wind speed and area-averaged cloudiness are shown to be negatively correlated with day-to-day lightning frequency over land, and thunderstorm occurrence is suppressed windward of, and enhanced leeward of mountain ranges. Lightning and environmental conditions are similarly related in the MJO. During break periods, the regular diurnal cycle of lightning is enhanced where ambient low-level winds are easterly but abnormally weak—in the Strait of Malacca, over western and southern Borneo and the adjacent seas, and in the region of nocturnal thunderstorms to the west of Sumatra and Java. When the active, cloudy phase of the MJO, accompanied by low-level westerly winds, passes over the Maritime Continent, lightning is enhanced leeward (to the east) of the mountains of Java, Borneo, and the Malay Peninsula. The most pronounced MJO-related lightning anomalies
are observed during the evening. The spatial patterns of lightning and rainfall anomalies are broadly similar, but lightning anomalies tend to be more concentrated near coastlines.

Section 4.1. Introduction

The term “Maritime Continent” is used in meteorological literature to denote to the region between southeastern Asia and Australia that contains numerous islands, including three of the six largest non-continental landmasses in the world (New Guinea, Borneo, and Sumatra; see Fig. 4.1a). The diurnal variability of convection over the Maritime Continent has been described based on satellite infrared imagery (Mori et al. 2004; Sakurai et al. 2005), satellite-based rainfall data from the Tropical Rainfall Measuring Mission (TRMM; Nesbitt and Zipser 2003; Mori et al. 2004; Kikuchi and Wang 2008; Wu et al. 2009; Fujita et al. 2010; Teo et al. 2011; Oh et al. 2012), lightning observations from the World Wide Lightning Location Network (WWLLN; Virts et al. 2012), and model simulations (Wu et al. 2009; Fujita et al. 2010; Teo et al. 2011). On sunny days, the land tends to be warmer by virtue of its smaller heat capacity. The stronger land-sea contrast gives rise to seaward pressure gradients along the coasts of the islands, which drive sea breezes that propagate inland during the afternoon. Convection firing along the sea breeze fronts produces rainfall and lightning in coastal areas. Solar heating also causes warm air to rise along the mountain slopes (Fig. 4.1a), producing convection during the late afternoon and evening. Figure 4.1c, d (after Teo et al. 2011) show the climatological diurnal evolution of lightning over one transect across Sumatra and the Malay Peninsula and another across Borneo. In both sections, afternoon and evening lightning is clearly visible over the coastlines and mountain slopes. At night, the local circulation patterns reverse. Cooler air subsides down the mountain slopes, producing convergence and convection over the adjacent lowlands during the
early morning. For example, in the valley between the mountain ranges on the Malay Peninsula in the transect (Fig. 4.1c, e), the enhancement of the lightning begins around 1600 LT, an hour or two later than over the mountains, and it continues into the evening, long after convection over the mountain ranges has ceased. Land breezes from Sumatra and the Malay Peninsula converge in the Strait of Malacca, producing frequent thunderstorms during the night and morning hours (Fig. 4.1c). In addition, gravity waves, as suggested by the numerical simulations of Mapes et al. (2003), appear to propagate away from the regions of afternoon convection and touch off nocturnal convection along the coastlines of the major islands, which subsequently propagates out to sea, as indicated by the black arrows in Fig. 4.1c, d. The annual-mean distribution of lightning over the Maritime Continent (Fig. 4.1b) reflects the mechanisms that govern its diurnal cycle: maxima are observed over coastlines, mountain slopes, and adjacent coastal waters and valley floors.

The climatological-mean pattern of convective variability is modulated by transient variability. For example, convective rainfall over the Maritime Continent is comparatively enhanced during La Niña years and suppressed during El Niño years (Ropelewski and Halpert 1987), as the rising branch of the Walker Circulation shifts between the Maritime Continent and the central Pacific. Lightning, in contrast, tends to be more frequent during El Niño years, particularly over western Borneo and Java and to the west of Sumatra (Hamid et al. 2001; Yoshida et al. 2007). In contrast, rainfall and lightning frequency of occurrence vary in phase with one another in association with the climatological-mean annual cycle (Chang et al. 2005; Virts et al. 2011). To lay the groundwork for understanding these complex interrelationships, in Section 4.3, we examine the impact of day-to-day variations in cloudiness and low-level wind on
the spatial distribution of thunderstorm formation and propagation over the Maritime Continent, using the datasets described in Section 4.2.

The Madden-Julian Oscillation (MJO) dominates atmospheric variability over the Maritime Continent at intraseasonal (~30-80 day) time scales. During a typical cycle of the MJO, an area of enhanced cloudiness and precipitation, accompanied by low-level convergence, develops over the central equatorial Indian Ocean, propagates eastward across the Maritime Continent, and dissipates in the western or central equatorial Pacific (Zhang 2005). Analyses of data from the Lightning Imaging Sensor (LIS), which is carried aboard the TRMM satellite, have shown that lightning tends to be enhanced during the “break” period of MJO precipitation (Morita et al. 2006), particularly over the islands of the Maritime Continent (Kodama et al. 2006). Analyzing WWLLN data, Virts et al. (2011) also noted more frequent lightning over the ocean to the west of Sumatra during the MJO break period. The modulation of the diurnal cycle of lightning over the Maritime Continent by the MJO has not yet been systematically investigated. In Section 4.4, we examine the spatial and temporal variability of lightning during both MJO break and active periods and the environmental conditions that drive these variations. Conclusions are presented in Section 4.5.

Section 4.2. Data

WWLLN (http://wwlln.net) is a ground-based network consisting of 68 sensors (as of October 2012) that monitor very low frequency (VLF) radio waves for lightning sferics. The network uses a time of group arrival technique (Dowden et al. 2002) on the detected sferic waveforms to locate lightning to within ~5 km and < 10 µs (Abarca et al. 2010). The global detection efficiency of the network is estimated to be ~10% of all strokes with preferential
detection of cloud to ground strokes (Rodger et al. 2009; Abarca et al. 2010). This detection efficiency allows the network to detect almost all lightning-producing storms (Jacobson et al. 2006). For this study, WWLLN observations over the Maritime Continent from 2008-2011 are averaged onto an hourly, $0.25^\circ \times 0.25^\circ$ grid to create a field of lightning frequency of occurrence.

The TRMM satellite was launched in 1997 into a $35^\circ$ inclination orbit carrying a Precipitation Radar (PR) and Visible and Infrared Scanner (VIRS). The TRMM 3B42 dataset contains TRMM rainfall observations supplemented with data from other satellite-borne microwave imagers and infrared sensors (e.g., Huffman et al. 2007) and is available at 3-hourly temporal resolution and $0.25^\circ \times 0.25^\circ$ spatial resolution.

This study also makes use of 850-hPa wind fields from the European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA)-Interim (Dee et al. 2011), which are available four times daily at $1^\circ$ horizontal resolution, and cloud observations from Clouds and Earth’s Radiant Energy System (CERES; Wielicki et al. 1996). The CERES_SYN dataset contains cloud area fractions observed by the Moderate Resolution Imaging Spectroradiometer (MODIS) instruments aboard the polar-orbiting Aqua and Terra satellites (with equatorial crossings at 1:30 LT and 10:30 LT, respectively) that are supplemented by observations from geostationary satellites to create daily, $1^\circ \times 1^\circ$ cloud fields (further details are at ceres.larc.nasa.gov).

The present study focuses on the day-to-day (Section 4.3) and intraseasonal (Section 4.4) modulation of lightning frequency. To highlight this variability, 15-day or 80-day high-pass Lanczos filters are applied to the gridded datasets as indicated. When this operation is performed, the data are referred to as “15-day hp filtered” and “80-day hp filtered,” respectively.
In both cases, the filtering removes the variability associated with the climatological-mean annual cycle, ENSO, and trends in the detection efficiency of WWLLN.

Section 4.3. Lightning variability: Impact of clouds and low-level winds

Lightning variability over the Maritime Continent associated with the annual cycle or MJO has sometimes been represented in terms of changes to the lightning frequency averaged over the entire region, without reference to the spatial distribution (e.g., Morita et al. 2006). Daily, 80-day hp filtered lightning at each grid point regressed onto 80-day hp filtered lightning averaged over the Maritime Continent (12°S to 8°N; 94°E to 159°E), shown in Fig. 4.2a, indicates that the domain-mean lightning frequency is dominated by lightning variability in the vicinity of the major islands (particularly the western islands of Sumatra, Borneo, and Java), where climatological-mean lightning is most frequent (Fig. 4.1b). On intraseasonal time scales, lightning in the Strait of Malacca and to the south and west of Sumatra, Borneo, and Java tends to vary in synchrony, as evidenced by the structure of the empirical orthogonal function (EOF) shown in Fig. 4.2b and verified by inspection of one-point correlation maps for these regions (not shown). Similar relationships are observed for 15-day hp filtered data (Fig. 4.2c).

Because thunderstorms over the Maritime Continent are often driven by local, diurnally-varying land-sea and mountain-valley wind regimes (see Section 4.1; Virts et al. 2012), it follows that environmental conditions that modulate the strength of these wind regimes should likewise modulate lightning frequency. Numerical simulations of atmospheric circulation in the presence of coastlines and topography indicate that stronger vertical velocities occur in association with stronger heating of the land surface (Qian et al. 2012b) on days with less extensive cloud cover (Segal et al. 1986). This relationship can be demonstrated by regressing
15-day hp filtered lightning at each individual grid point onto 15-day hp filtered domain-mean cloudiness, as shown in Fig. 4.2d. When cloudy conditions prevail over the Maritime Continent, lightning tends to be suppressed over land, consistent with reduced solar heating leading to weaker diurnally-varying circulations.

The strength of land-sea and mountain-valley wind regimes is also modulated by the ambient, or background, winds (Miller et al. 2003, and references therein; Qian et al. 2010, 2012a). Point-wise regression maps of 15-day hp filtered lightning (Fig. 4.3a) and rainfall (Fig. 4.3b) onto 850-hPa scalar wind speed show a marked land/sea contrast, with negative values (i.e., less rain and lightning observed in association with stronger ambient winds) over land and positive values over the seas. We conclude from this result that strong ambient low-level winds disrupt the local land-sea and mountain-valley wind regimes, leading to less vigorous convection and reduced incidence of thunderstorms over land, in agreement with the observations of Fujita et al. (2011) and Qian et al. (2012a).

Wind direction relative to the terrain also modulates the location and frequency of occurrence of lightning, as shown in Fig. 4.4. For each of the three major islands, a time-varying index of the 850-hPa wind component perpendicular to the crest of the mountain range has been generated based on an average of the grid points that lie along the black line. 15-day hp filtered 850-hPa wind, lightning, and precipitation fields were then regressed onto the wind index for each island. Striking contrasts are observed in each case, with lightning and precipitation suppressed windward of the mountains and enhanced in the lee. Previous studies have also noted this relationship between precipitation and low-level wind: TRMM precipitation in the vicinity of Borneo and New Guinea composited according to synoptic low-level wind regimes shows enhanced precipitation in the lee of the mountains (Ichikawa and Yasunari 2006, 2008); Sakurai
et al. (2005) analyzed cloud-top temperatures over Sumatra and found lee propagation in cases with low-level westerly winds.

The development of lee precipitation has also been investigated in numerical simulations (e.g., Sato and Kimura 2003; Mori et al. 2004; Sasaki et al. 2004; Wu et al. 2009). The simulations suggest that breezes blowing up valleys converge and produce a layer of moist air and convection along the length of a mountain range during the afternoon. Ambient winds above the crests then advect the moist air to the lee of the mountains, and cold outflow and gravity waves from the daytime convection provide the lift required to trigger convection that then propagates farther leeward. Confirmation for this explanation is found in Fig. 4.5, in which the lightning regression patterns in Fig. 4.4 are divided into three segments by hour of the day. The dipole patterns develop during the afternoon hours, most prominently over New Guinea, and propagate away from the mountain ranges, onto the adjacent lowlands during the evening.

In this section, we have demonstrated that environmental conditions such as cloudiness and ambient wind speed and direction that affect the strength and evolution of the local diurnally-varying land-sea and mountain-valley wind regimes also modulate thunderstorm occurrence on a day-to-day basis. In the following section, we turn to the particular case of the modulation of lightning by the Madden-Julian Oscillation, which is associated with intraseasonal variations in both circulation and cloudiness over the Maritime Continent.

Section 4.4. Lightning variability: Impact of MJO

The time-varying state of the MJO can be represented by the Real-Time Multivariate MJO (RMM) index, which is calculated by projecting daily observations onto the first two EOFs of a multivariate field made up of near-equatorial outgoing longwave radiation (OLR) and lower-
(850 hPa) and upper-tropospheric (200 hPa) zonal wind (Wheeler and Hendon 2004). The evolution and strength of the MJO is frequently represented by composite maps of atmospheric variables associated with eight linear combinations of the two RMM indices (e.g., Wheeler and Hendon 2004). In this study, each day of the observation period (2008-2011 for WWLLN and ERA-Interim; 2007-2010 for CERES; 1998-2011 for TRMM) was assigned to whichever of the eight phases that it projects onto most strongly; days on which the magnitude of the MJO vector was less than one standard deviation from zero were discarded.

The mean cloud fraction, precipitation rate, and lightning frequency across the Maritime Continent during each MJO phase are shown in Fig. 4.6 (adapted from Fig. 12 of Morita et al. 2006), and maps of each variable regressed onto time series representing MJO phases 1-8 are shown in Fig. 4.7. Patterns for phases 5-8 are identical to those for phases 1-4 but with sign reversed. Cloudiness and precipitation increase across the domain as the region of active convection approaches from the west and then decrease as the region of active convection shifts eastward into the western Pacific. In contrast, domain-mean lightning is least frequent around the time of peak cloudiness and precipitation (Fig. 4.6). The spatial pattern of MJO lightning anomalies is visually complex (Fig. 4.7c). Composites of CERES cloud fraction and ERA-Interim 850-hPa wind during MJO phases 8-1-2 and 4-5-6 are shown in Fig. 4.8 and 4.9. Analogous plots based on OLR are shown in Rauniyar and Walsh (2011). Extensive cloudiness is observed over the major islands during all MJO phases, but the difference plots in Fig. 4.8c and 4.9c illustrate the overall cloudier conditions and greater prevalence of westerly winds during phases 4-5-6. Accordingly, the composites presented in the remainder of this section were generated for days during phases 8-1-2 (which we refer to as the MJO “break” period) and 4-5-6 (the “active” period).
Composites of daily TRMM precipitation and WWLLN lightning during MJO phases 8-1-2 and 4-5-6 are shown in Fig. 4.10 and 4.11, respectively. Lightning, an indicator of the most vigorous convection, tends to be concentrated near the coasts of the large islands during both periods, while large precipitation anomalies are observed over both land and sea. The difference plot shown in Fig. 4.11c reveals contrasting lightning anomalies across the major landmasses of the Maritime Continent—for example, western and southern Borneo experience more lightning during the MJO break period, while eastern Borneo experiences more lightning during the active period. Contrasts are also observed across Sumatra, Java, the Malay Peninsula, and the Celebes. Lightning is more frequent over most of New Guinea during break periods. Contrasts in rainfall rates shown in Fig. 4.10c (see also Rauniyar and Walsh [2011], based on TRMM data) are also observed across Sumatra, Borneo, and to a lesser extent the Celebes in association with the MJO, with more rain falling to the west of the islands during the break period. Most of the seas surrounding the islands receive more rain during the active MJO period, in agreement with results of Wu and Hsu (2009) and Oh et al. (2012), both based on TRMM data.

As described in Section 4.1 and in Virts et al. (2012; their Fig. 6), the diurnal cycle of lightning over the Maritime Continent can be summarized as follows: afternoon thunderstorms are most frequent over the smaller islands and along the coastlines of the larger islands; evening thunderstorms are observed over the mountain slopes and near-coastal waters of the larger islands; and morning thunderstorms are most frequent over the seas surrounding the islands and over lowlands adjacent to mountain ranges. Figures 4.12 and 4.13 show the difference in precipitation rate and lightning frequency between MJO phases 8-1-2 and 4-5-6 during these three segments of the day, and Fig. 4.14 shows the hourly difference in lightning between break
and active periods over the same transects as Fig. 4.1c, d. Similar results (not shown) are obtained when compositing is performed separately for 2008-2009 and 2010-2011. Afternoon precipitation and thunderstorms associated with sea breeze fronts tend to be more frequent over the coastal areas of Sumatra and Borneo during the MJO break period. The largest lightning anomalies, however, occur during evening. During the MJO break period, lightning and precipitation are enhanced over the western and southern mountain slopes and coastal areas of Sumatra, Java, the Malay Peninsula, Borneo, and the Celebes (Fig. 4.12b, 4.13b), and the offshore propagation of the thunderstorms continues into the morning hours (Fig. 4.13c, 4.14).

The low-level winds during the break phase exhibit an easterly component but are weak over the major islands (Fig. 4.9a). Previous studies have noted the development and leeward propagation of long-lived mesoscale convective systems (MCSs) over Borneo under weak low-level easterlies (Ichikawa and Yasunari 2006) and the correspondence between weak background winds and enhanced morning precipitation in the Strait of Malacca (Fujita et al. 2010), over western Borneo (Wu et al. 2008), and to the west of Sumatra (Fujita et al. 2011).

As the active phase of the MJO moves over the Maritime Continent, the low-level winds acquire a westerly component and strengthen (Fig. 4.9b). Afternoon lightning tends to be suppressed over land, where the atmosphere is more stable than during the break period (Rauniyar and Walsh 2011), but more evening thunderstorms are observed over the northern and eastern mountain slopes and coastal areas of the major islands (Fig. 4.13, 4.14). These and other storms touched off by gravity waves propagate offshore during the morning hours (Fig. 4.14), resulting in enhanced lightning and precipitation over areas such as the Java Sea during the active phase (Rauniyar and Walsh 2011; Oh et al. 2012). The exception to this pattern is New Guinea, which experiences relatively lighter winds during MJO phases 4-5-6 compared to 8-1-2
but experiences enhanced lightning over its mountains and near-coastal waters during phases 8-1-2.

The composites shown thus far suggest that the stronger diurnal heating of land surfaces during the less-cloudy MJO break period could lead to more intense convection and more lightning, a mechanism previously suggested by Sui and Lau (1992) and Kodama et al. (2006). To test the validity of this mechanism and extend it to wind, we attempt to predict MJO-related anomalous lightning frequency using the relationships between lightning and low-level winds and cloudiness documented in Section 4.3. At each grid point, for each MJO phase, the following linear prediction scheme for the anomalous lightning frequency ($L$) associated with one standard deviation of the MJO index is used:

$$L = a_{L,CF} b_{CF,MJO} + a_{L,u} b_{u,MJO} + a_{L,v} b_{v,MJO},$$

where $CF$ with an overbar is the domain-mean cloud fraction, $u$ and $v$ are components of the 850-hPa wind at each grid point, $a$ is the regression coefficient of 15-day hp filtered lightning onto the indicated 15-day hp filtered variable, and $b$ is the regression coefficient of the indicated 80-day hp filtered variable onto the MJO index for that phase. This linear regression analysis generates a predicted map of lightning frequency for each MJO phase based solely on the day-to-day relationships between lightning and selected environmental factors. Use of 15-day hp filtered data in estimating the $a$ coefficients ensures that they are not influenced by the MJO.

Maps of observed and predicted lightning for MJO phases 8-1-2 minus 4-5-6 are shown in Fig. 4.15a, b, respectively. The observed maps were generated by compositing maps of 80-day hp filtered lightning frequency regressed onto these MJO phases. The predicted lightning field captures the enhanced lightning west of Sumatra and over the Strait of Malacca and western Borneo during the break period and over the South China and Java Seas during the active period.
The observed minus predicted difference plot shown in Fig. 4.15c illustrates the areas where the prediction fails to capture the observed lightning pattern, most notably over eastern Borneo, where the residual lightning is as large or larger than the observed. Clearly, a much more sophisticated model would be needed in order to accurately predict lightning variability associated with the MJO. Nevertheless, the similarities between the observed and predicted lightning fields in Fig. 4.15 demonstrate that, to first order, MJO-related lightning variability over the Maritime Continent during the active and break periods reflects the modulation of the local, diurnally-varying wind regimes by the large-scale cloud and circulation anomalies that accompany the MJO.

MJO modulation of the strength of the diurnal cycle of lightning is illustrated by the difference map, shown in Fig. 4.16, between the amplitude of the first harmonic of hourly-mean lightning frequency during MJO phases 8-1-2 and 4-5-6. The strong similarity between Fig. 4.11c and 4.16 suggests that the complex spatial structure in the difference field of lightning frequency between MJO active and break periods reflects the differences in the strength of the diurnal cycle of lightning frequency; i.e., that the MJO mediates lightning over the Maritime Continent mainly through its influence on the diurnal cycle.

Our analysis of MJO-related diurnal lightning variability has focused on the spatial pattern and amplitude of its diurnal cycle, as represented by lightning frequency during fixed segments of the day (Fig. 4.13). Analysis of phase of the diurnal cycle of lightning during the MJO cycle, which shows distinctive shifts in the timing of the peak lightning, is presented in Appendix A4.

**Section 4.5. Conclusions**
Over the Maritime Continent, land-sea breezes and mountain-valley wind regimes forced by the diurnal cycle in low-level heating are the dominant mechanism for producing the strong lifting required to generate thunderstorms. We have demonstrated that lightning over the islands is less frequent when these local regimes are disrupted by strong ambient low-level winds (Fig. 4.3) or are weakened by a decrease in low-level heating due to increased cloudiness (Fig. 4.2d). Anomalous low-level wind components perpendicular to the mountain ranges also influence the diurnal cycle in convection: lightning and precipitation are suppressed windward of the mountains, and lightning is enhanced over the lee mountain slopes during the day and the adjacent lowlands during the evening and morning (Fig. 4.4, 4.5).

The response of lightning to changes in cloudiness and low-level wind is observed not only on a day-to-day basis (Section 4.3) but also in association with the intraseasonal MJO. During the MJO break period, a time of generally suppressed cloudiness (Fig. 4.8a) and weak, easterly low-level winds (Fig. 4.9a), thunderstorms develop more frequently over the western mountain slopes and coastlines of the major islands and propagate westward over the seas during the night and morning (Fig. 4.11, 4.12). In contrast, the enhanced cloudiness (Fig. 4.8b) and westerly winds (Fig. 4.9b) observed during the MJO active period are associated with more frequent thunderstorm development over the eastern slopes and coastlines and adjacent areas, including the Java Sea (Fig. 4.11). During both active and break periods, the MJO-related wind and cloudiness anomalies serve to amplify the climatological-mean diurnal cycle in some areas and weaken it in others.

Cloudiness and lightning over the Maritime Continent also vary inversely in association with the ENSO cycle (see Section 4.1). Although our data period includes an El Niño year, 2009, and a La Niña year, 2010, a comparison of TRMM diurnal precipitation anomalies over
the Maritime Continent during these years with ENSO anomalies based on the longer TRMM data record (not shown) indicates that 2009-2010 anomalies are not representative of typical ENSO variability. Yoshida et al. (2007) composited LIS lightning over an interval encompassing the El Niño years of 1998 and 2002 and the La Niña year of 1999. The spatial patterns of ENSO-related lightning in their Fig. 4.2 and of MJO-related lightning in our Fig. 4.11 are similar in many respects. Enhanced lightning over and to the west of Sumatra and western Borneo is observed during El Niño years and during MJO phases 8-1-2. This suggests that the reduced cloud cover over the Maritime Continent may strengthen the diurnally-varying local circulations in the same manner during El Niño years as during MJO break periods. Low-level winds over the western Maritime Continent are anomalously westerly during La Niña years (Yoshida et al. 2007) and during MJO active phases, and in both cases lightning is enhanced over eastern Borneo and to the east of the Malay Peninsula, i.e., on the lee side of the nearby mountains relative to the anomalous winds. The relationship between rainfall and downslope winds in the lee of the mountains of Borneo holds for both El Niño and La Niña years (Qian et al. 2012a, based on NOAA Climate Prediction Center-CPC Morphing Technique [CMORPH] data).

The analysis of the dependence of lightning on day-to-day fluctuations in cloudiness and low-level wind (Section 4.3), the usefulness of these empirical relationships for predicting the spatial pattern of lightning anomalies during the active and break periods of the MJO (Section 4.4, particularly Fig. 4.15), and the similarities between these results and the previously published composites of lightning and wind associated with ENSO (Yoshida et al. 2007), confirm that thunderstorm variability over the Maritime Continent is mediated by variations in
the strength of the local land/sea and mountain/valley wind regimes in response to changes in ambient atmospheric conditions.

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Lightning location data were provided by WWLLN (http://wwlln.net), a collaboration among over 50 universities and institutions. TRMM rainfall data were obtained from NASA’s Goddard Earth Sciences Data and Information Services Center (http://disc.sci.gsfc.nasa.gov/), ERA-Interim wind data from the European Centre for Medium-Range Weather Forecasts (http://ecmwf.int/), and CERES cloud data from NASA’s Langley Research Center (http://ceres.larc.nasa.gov/).
Figure 4.1. Elevation ($a$; m) and climatological WWLLN frequency of occurrence of lightning ($b$; strokes km$^{-2}$ yr$^{-1}$) for the Maritime Continent. Climatological diurnal cycle (repeated for clarity) of lightning along black lines in top panel ($c$, $d$; local time taken from nearby cities indicated by black stars). In this and all subsequent figures, the color scale is designed to capture
the most interesting features and may be saturated at one or both ends. Elevation (m) along cross-sections is in panels e, f.
Figure 4.2. 80-day hp filtered WWLLN lightning (strokes km$^{-2}$ yr$^{-1}$) regressed onto 80-day hp filtered time series of lightning averaged over the entire domain (a). 80-day hp filtered, 0.25° × 0.25° WWLLN lightning regressed onto PC1 of 80-day hp filtered, 1° × 1° lightning (b left; analogous results for 15-day hp filtered lightning are in c). The variance explained is indicated
in the lower right corner. 15-day hp filtered lightning regressed onto 15-day hp filtered time series of CERES-MODIS cloud area fraction \( (d) \) averaged over the entire domain. Black contours indicate the 500-m elevation.
Figure 4.3. 15-day hp filtered WWLLN lightning (top; strokes km$^{-2}$ yr$^{-1}$) and TRMM precipitation (bottom; mm day$^{-1}$) regressed point-wise onto 15-day hp filtered ERA-Interim scalar wind speed (m s$^{-1}$; $1^\circ \times 1^\circ$ resolution). Black contours indicate the 500-m elevation.
Figure 4.4. 15-day hp filtered 850-hPa ERA-Interim winds (a; m s⁻¹), TRMM precipitation (b; mm day⁻¹), and WWLLN lightning (c; strokes km⁻² yr⁻¹) regressed onto 15-day hp filtered time series of component of 850-hPa wind perpendicular to the black line. Black contours indicate the 500-m elevation.
Figure 4.5. 15-day hp filtered WWLLN lightning (strokes km⁻² yr⁻¹) during indicated segments of the day regressed onto 15-day hp filtered time series of component of 850-hPa wind perpendicular to the black line. Local times calculated separately for each island. Black contours indicate the 500-m elevation.
Figure 4.6. CERES-MODIS cloud area fraction (a), WWLLN lightning (b, dashed line with scale at left), and TRMM precipitation (c, solid line with scale at right) during each MJO phase, averaged over the Maritime Continent (12°S to 8°N, 94°E to 159°E).
Figure 4.7. 80-day hp filtered CERES-MODIS cloud (left), TRMM precipitation (center), and WWLLN lightning (right) regressed onto time series representing MJO phases 1-8. Regression coefficients are scaled by the annual mean. Adapted from Virts et al. (2011).
Figure 4.8. Mean CERES-MODIS cloud area during MJO phases 8-1-2 (a) and 4-5-6 (b), and difference between them (c). Black contours indicate the 500-m elevation.
Figure 4.9. As in Fig. 4.8, but for ERA-Interim 850-hPa winds (velocity, in m s\(^{-1}\), indicated by colored shading). Black contours indicate the 500-m elevation.
Figure 4.10. As in Fig. 4.8, but for TRMM precipitation (mm day$^{-1}$).
Figure 4.11. Mean WWLLN lightning (strokes km^{-2} yr^{-1}) during MJO phases 8-1-2 (a) and 4-5-6 (b), and difference between them (c). Black contours indicate the 500-m elevation.
Figure 4.12. Difference between TRMM precipitation (mm day$^{-1}$) during MJO phases 8-1-2 and 4-5-6, for the indicated segments of the day. Times are given in Singapore LT. Black contours indicate the 500-m elevation.
Figure 4.13. Difference between WWLLN lightning (strokes km$^{-2}$ yr$^{-1}$) during MJO phases 8-1-2 and 4-5-6, for the indicated segments of the day. Times are given in Singapore LT. Black contours indicate the 500-m elevation.
Figure 4.14. As in Fig. 4.1, but for difference between lightning during MJO phases 8-1-2 and 4-5-6. Black arrows are repeated from Fig. 4.1.
Figure 4.15. Difference between 80-day hp filtered WWLLN lightning regressed onto MJO phases 8-1-2 and 4-5-6 (top). Difference between lightning for MJO phases 8-1-2 and 4-5-6, as predicted by regressions onto ERA-Interim 850-hPa winds and CERES-MODIS cloudiness (middle; see text for details). Difference between observed and predicted lightning (bottom; i.e., top panel minus middle panel). Black contours indicate the 500-m elevation.
Figure 4.16. The difference between the amplitude of the first harmonic of hourly-mean lightning frequency (from WWLLN; strokes km$^2$ yr$^{-1}$) during MJO phases 8-1-2 and 4-5-6. Black contours indicate the 500-m elevation.
Appendix to Chapter Four

Section A4.1. Lightning diurnal phase: Impact of MJO

In order to investigate shifts in the timing of the diurnal cycle of lightning during the MJO, harmonic analysis has been applied to hourly-mean lightning frequency at each grid point during MJO phases 8-1-2 and 4-5-6. The hour of the maximum of the first harmonic, shown in Fig. A4.1a, b, is consistent with an afternoon and evening peak in lightning over land, offshore propagation during the night, and maximum morning lightning over the seas, in agreement with the diurnal summary in Section 4.1. Previous studies of shifts in the diurnal cycle of convection during the MJO have shown that while no shift is observed in the diurnal phase of deep convective cloud amounts averaged over the Maritime Continent region (Tian et al. 2006), the diurnal maximum in TRMM rainfall occurs 1-3 hours earlier during the break (Rauniyar and Walsh 2011) and developing stages (i.e., phase 3) of the MJO than during the active period, particularly over the ocean (Oh et al. 2012). The difference plot in Fig. A4.1c indicates that, with some exceptions, areas with more lightning during the break period, such as the seas southwest of Sumatra and Borneo and west of the Celebes, tend to experience an earlier lightning peak during the break period than during the active period (i.e., they exhibit negative values in Fig. A4.1c), in agreement with Rauniyar and Walsh (2011) and Oh et al. (2012). Anomalies in the vicinity of land, where there is sufficient lightning for harmonic analysis, range up to 2-3 hours. The Java Sea and seas east of the Celebes and the Malay Peninsula, which exhibit more lightning during the active period, tend to experience an earlier lightning peak during the active period than during the break period.
Figure A4.1. Hour of the maximum of the first harmonic of hourly-mean lightning frequency (from WWLLN; flashes km$^{-2}$ yr$^{-1}$; Singapore LT) during MJO phases 8-1-2 ($a$) and 4-5-6 ($b$), and the difference between them ($c$). Black contours indicate the 500-m elevation. Gray shading indicates areas of low annual-mean lightning or weak diurnal lightning variability.
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