An atmospheric investigation of California's August 2020 lightning-initiated wildfire events

Sydney Lybrand

Follow this and additional works at: https://louis.uah.edu/uah-theses

Recommended Citation
https://louis.uah.edu/uah-theses/456

This Thesis is brought to you for free and open access by the UAH Electronic Theses and Dissertations at LOUIS. It has been accepted for inclusion in Theses by an authorized administrator of LOUIS.
AN ATMOSPHERIC INVESTIGATION OF CALIFORNIA’S AUGUST 2020 LIGHTNING-INITIATED WILDFIRE EVENTS

Sydney Lybrand

A THESIS

Submitted in partial fulfillment of the requirements for the degree of Master of Science in Atmospheric Science in The Department of Atmospheric and Earth Science to The Graduate School of The University of Alabama in Huntsville

May 2023

Approved by:
Dr. Chris Schultz, Research Advisor
Dr. John Mecikalski, Committee Chair
Dr. Larry Carey, Committee Member
Dr. John Mecikalski, Department Chair
Dr. Rainer Steinwandt, College Dean
Dr. Jon Hakkila, Graduate Dean
Abstract

AN ATMOSPHERIC INVESTIGATION OF CALIFORNIA’S AUGUST 2020 LIGHTNING-INITIATED WILDFIRE EVENTS

Sydney Lybrand

A thesis submitted in partial fulfillment of the requirements for the degree of Master of Science in Atmospheric Science

Atmospheric and Earth Science
The University of Alabama in Huntsville
May 2023

This study examines two events from August 2020 in California: August 16 and August 24. Both days were forecast to have large lightning-initiated wildfire potential. Herein, environmental, satellite, radar, and lightning characteristics of the atmosphere have been assessed to determine why one event was such a prolific producer of lightning-initiated wildfire and the other was not. Notable differences in the vertical structures of thunderstorms were present, particularly in the mixed-phase region where shallower mixed-phase depths and smaller ice masses were present on August 24th. This study suggests that it is physically plausible that aerosol load from prior fires contributed to weaker mixed-phase regions in thunderstorms; however, some thermodynamic differences, particularly in convective available potential energy and deep moisture, were present and must be considered in addition to aerosol information. This study suggests there is potential utility in considering aerosol concentration and depth while forecasting fire potential.
Acknowledgements

Thank you to NASA Short-Term Prediction Research and Transition Center (SPoRT) Applied Research Award Number 001949-00032 for supporting this work. I would like to acknowledge my advisor Dr. Schultz for being an amazing advisor, friend, and life coach. I greatly appreciate you and your time. I am grateful for my committee members Dr. Mecikalski and Dr. Carey and their time and advice. Y’all are awesome. I am thankful for my NASA SPoRT family, especially Mr. Wade, Mr. Meyer, and Mr. Allen. I also want to acknowledge the NOAA/NESDIS Center for Satellite Applications and Research for the assistance in viewing and analyzing aerosol optical depth data. I am grateful for my family, especially my Mama and Daddy. Thank you for loving me and caring about me even when I don’t deserve it. I have the best support system a girl could ask for. Lastly, thank you to God for creating me and loving me.
# Table of Contents

Abstract ................................................................. ii

Acknowledgements ......................................................... iv

Table of Contents ............................................................. vii

List of Figures ............................................................... viii

List of Tables ................................................................. xi

Chapter 1. Introduction and Overview ................................. 1

Chapter 2. Background ....................................................... 4

2.1 Relationship between Convection, Aerosols, and Lightning Potential ................................................. 5

2.2 Aerosols and Cloud Formation ........................................ 7

2.2.1 The Direct Effect ......................................................... 8

2.2.2 The Indirect Effect ......................................................... 9

2.2.3 The Semi-direct Effect .................................................... 11

2.3 Charge Separation and Lightning Formation .................... 12

2.3.1 Selective Ion Capture ..................................................... 12
2.3.2 Inductive Charging ............................................. 13
2.3.3 Non-inductive Charging ................................. 13
2.4 Charging and Wildfire Relationship ................. 18

Chapter 3. Data and Methods ................................. 19
3.1 Atmospheric Forecast Models ......................... 19
3.2 Surface Observations and Atmospheric Profiles .... 20
3.3 Radar ................................................................. 21
3.4 Satellite Data ....................................................... 24
3.5 Lightning Data ..................................................... 25
3.6 Study Area and Grid Formation ....................... 26
3.7 Storm Tracking ..................................................... 27

Chapter 4. Results .................................................. 30
4.1 Land Conditions .................................................. 30
4.2 Synoptic Pattern .................................................. 32
   4.2.1 500 mb ....................................................... 33
   4.2.2 850 mb ....................................................... 38
   4.2.3 1000 mb ..................................................... 43
4.3 Thermodynamic Influences ............................. 49
# List of Figures

1.1 NIFC forecasted risk of significant fire potential for August 16. . . 2  
1.2 NIFC forecasted risk of significant fire potential for August 24. . . 2  

2.1 Diagram showing thunderstorm evolution and electric field formation processes. ................................................. 7  
2.2 Diagram showing aerosol impacts and the ways they can alter the cloud. ................................................................. 8  
2.3 Schematic showing the impact of polluted air on cloud nuclei concentration. .............................................................. 10  
2.4 Schematic showing inferred mass and charge transfer mechanisms between graupel and ice. ........................................ 14  
2.5 Conceptual model of charging in a thunderstorm and the development of cloud charging layers. .............................. 16  
2.6 Diagram showing layer net charge and evolution through time as lightning occurs. ..................................................... 17  

3.1 Study region showing the KMUX and KBBX radar location as well as the OAK upper air sounding site. ............................... 27  

4.1 U.S. monthly drought outlook for August 2020. ....................... 31  
4.2 Fuel moisture time series for the San Jose, CA RAWS site. ....... 32  
4.3 Geopotential height patterns for 0000 UTC on August 16 and August 24. ................................................................. 34  
4.4 Geopotential height patterns for 0200, 0600, and 0900 UTC on August 16 and August 24. .............................................. 35  
4.5 Geopotential height patterns for 1100, 1300, and 1500 UTC on August 16 and August 24. .............................................. 36
4.6 Geopotential height patterns for 1700, 1900, and 2200 UTC on August 16 and August 24. .................................................. 37
4.7 Geopotential height patterns for 2300 on August 16 and August 24. 38
4.8 Geopotential height patterns for 0400 and 0600 UTC on August 16 and August 24. .................................................. 39
4.9 Geopotential height patterns for 0800 and 1000 UTC on August 16 and August 24. .................................................. 41
4.10 Geopotential height patterns for 1200, 1400, 1600, and 1900 UTC on August 16 and August 24. .................................................. 42
4.11 Geopotential height patterns for 2100 and 2300 UTC on August 16 and August 24. .................................................. 44
4.12 Geopotential height patterns for 0000 and 0200 UTC on August 16 and August 24. .................................................. 45
4.13 Geopotential height patterns for 0600, 1000, and 1200 UTC on August 16 and August 24. .................................................. 46
4.14 Geopotential height patterns for 1500, 1700, and 2000 UTC on August 16 and August 24. .................................................. 47
4.15 Geopotential height patterns for 2300 UTC on August 16 and August 24. .................................................. 48
4.16 Representative sounding from the Oakland, CA upper air site for August 16 and 24, 2020. .................................................. 50
4.17 Cell cloud-to-ground NLDN counts per cell per minute compared to overall cell mass for each day. ................................. 59
4.18 The number of negative and positive CG lightning detected within each cell each day. .................................................. 60
4.19 The percentile statistical analysis for the key MRMS variables. .......................... 61
4.20 Echo top, composite reflectivity, freezing level, and -10°C reflectivity per cell as a function of ice mass. .......................... 62
4.21 The correlation matrices for the MRMS variables for August 16th and August 24th. .................................................. 64
4.22 VIIRS satellite overpass and Mt. Chaul webcam view showing a visual comparison of each day. ................................. 65
4.23 A visual depiction of aerosol optical depth for August 16 and August 24. .......................................................... 66
4.24 CALIPSO overpass showing vertical feature mask. ............... 68
4.25 CALIPSO overpass showing vertical feature mask. ............... 68
4.26 CALIPSO overpass showing vertical feature mask. ............... 69
4.27 US EPA air quality data for the San Jose site showing the yearly trend of PM 2.5 over the 2020 period. ................... 70

5.1 A conceptual model of the August 16th setup and common processes assumed to impact the day. ............................. 82
5.2 A conceptual model of the August 24th setup and common processes assumed to impact the day. ............................. 83
List of Tables

4.1 Thermodynamic parameters assessed at the regularly launched NWS sounding times for August 16th and 24th for Oakland, CA. . . . . 55
4.2 MRMS parameters for the days of interest. . . . . . . . . . . . 58

5.1 A table identifying common aerosol processes, relevant sources, and the pros and cons for lightning production in convective storms. . 78
Chapter 1. Introduction and Overview

From 1992 to 2012, lightning-initiated wildfires (LIW) were responsible for 56 percent of total acreage burned by wildfires within the United States, burning an average of 2.3 million acres per year (Balch et al. 2017). California’s 2020 wildfire season was record-breaking as the California Department of Forestry and Fire Protection (CAL FIRE) estimated a record 4,257,863 acres were burned, 10,488 structures damaged or destroyed, and loss of 33 lives (CAL FIRE - 2020 Incident Archive).

One of the more active periods of the 2020 wildfire season began on August 16. In the days leading up to this event, the National Interagency Fire Center’s (NIFC) forecasted hazards were for moderate to low risk of significant fire potential in California, seen in figure 1.1. An early morning dry lightning event quickly started a multitude of fires as this system produced frequent lightning and strong, gusty winds.

By noon, 22 fire starts were reported (CAL FIRE - 2020 Fire Siege Report). Some of these fire starts were in hard-to-access areas due to the terrain, which prolonged the fire-fighting process. Over time, the fires rapidly spread, producing 4 of the top 20 wildfires in California history. Models and forecast products showed a similar forecast for August 24th, 2020. The area’s NIFC hazards for
August 24th included a high-risk LIW potential based on model guidance for the
time period, seen in figure 1.2. August 24th did not perform as expected as there were zero LIW starts despite the forecast for an enhanced risk of LIW potential.

The motivation for this work is to determine the physical and environmental differences between conditions on August 16th and August 24th in order to improve future wildfire forecasts. Herein, this study uses historical model output and multi-sensor remote sensing to examine the synoptic and mesoscale conditions on August 16th and August 24th. The working hypotheses for this study are:

1.) Differences in mixed-phase storm-scale characteristics resulted in weaker storms and fewer lightning-initiated wildfire starts on 24 August 2020 as compared to 16 August 2020.

2.) Furthermore, enhanced concentrations of aerosols present during the 24 August 2020 event from ongoing wildfires in the region may be the leading factor in mitigating LIW potential on that day compared to 16 August 2020, despite similar forecasted environmental conditions.

If statistically significant differences are determined through the examination of both days, there could be positive impacts to the wildfire community if similar events are encountered in the future.
Wildfires are hazardous and threaten life and property in many ways. From the early days of fire weather research, lightning was one of the main contributors to wildfire occurrence and risk (Palmer 1917). During the next decade, the number of thunderstorm days, storm paths, and storm precipitation were tracked to understand wildfire risk from lightning (Gisborne 1931). The time of day, direction of storm movement, and surface characteristics immediately below the storm were vital parameters for wildfire assessment. Over the next few decades, not much changed about the understanding of the causes of lightning fires. In the late 1970’s, the United States Department of Agriculture Forest Service (USDA-FS) created a model estimating lightning fire ignition potential where the input parameters were lightning activity, storm movement, fuel moisture, and fuel bulk density (Fuquay et al. 1979).

During the 1980’s, technology was implemented to pinpoint locations of lightning strikes, the time they occur, their polarity, and estimate the strength of the lightning flash (Cummins and Murphy 2009, Murphy et al. 2021). This technology revolutionized wildfire detection, allowing researchers to understand the conditions present at the time of wildfire ignition. Empirical methods were developed to combine fuel moisture, precipitation, and lightning flash density to
estimate ignition efficiency and where fire starts may begin based on the forecast (Rorig and Ferguson 1999). Recent work has focused on temperature and previously burned areas (Holsinger et al. 2016); seasonal and diurnal variations (Dowdy and Mills 2012); dry lightning occurrence (Dowdy and Mills 2012, Sopko et al. 2016, Vant-Hull et al. 2018); polarity of cloud-to-ground lightning (Fuquay 1982, Schultz et al. 2019); and properties of present continuing current (Rust et al. 1985, Saba et al. 2006, Bitzer 2017, Fairman 2022).

When considering these cases and their outcomes, reviewing the ways in which a storm cell is thought to undergo charge separation and lightning formation is helpful. It is also important to understanding how microphysical changes in the cloud impact charge separation and lightning formation. It is necessary to analyze the current understanding of how aerosols are thought to impact cloud formation given these microphysical processes. Lightning potential is another parameter of interest due to the differences in aerosol coverage on the two days.

2.1 Relationship between Convection, Aerosols, and Lightning Potential

A general understanding of the present atmospheric characteristics is helpful in diagnosing convective storm potential and lightning production chances. For typical storm formation considering ingredients-based forecasting, prior studies have shown that lifting, instability, and moisture must be present for any deep, moist convection to occur (McNulty 1978, McNulty 1995, Doswell 1987, Johns and Doswell 1992). Cloud condensation nuclei (CCN) present in the air serve as
areas for condensation to occur as air reaches the lifting condensation level (LCL). Water vapor then begins condensing on the available CCN. Condensational latent heating then enhances already present instability through thermodynamic processes. This enhanced instability leads to atmospheric lifting as the parcel becomes warmer than the surrounding environmental air.

Available CCN are necessary for moist convective initiation due to this process. Many CCN are naturally occurring in the atmosphere from sources such as smoke, dust, or other atmospheric aerosols. The amount and density of the aerosol coverage can change the optical properties of clouds, the clouds themselves, and the temperature profiles of the atmosphere surrounding the clouds (Rosenfeld and Givati 2006, Ackerman et al. 2000, Koren et al. 2004, Chylek and Wong 1995). Dye et al. (1986) describe how a storm cell typically forms charging layers after that initial time where CCN were required for cloud droplet formation and subsequent convective cloud growth in figure 2.1. The convection creates turbulent motions within the cloud, eventually leading to the growth of the cloud. As the cloud reaches areas of cooler temperatures, these cooler temperatures support the formation of supercooled liquid water, ice, and graupel: all necessary ingredients for a mixed-phase region of the cloud. This mixed-phase area is important as these mixed phases are necessary for charge separation in moist convective thunderstorms. The electric field grows as more hydrometeors exchange mass and charges separate. By the time the cloud reaches the mature stage, the mixed-phase region is theoretically well-established and sufficient for charge separation.
2.2 Aerosols and Cloud Formation

Aerosols are thought to impact precipitation and cloud processes in many ways, whether it be looked at in an orographic respect (Rosenfeld and Givati 2006), from a microphysical perspective (Xue et al. 2010), or possibly through radiation changes (Takemura et al. 2005), among other ways. These and many other previous studies have examined the relationship between aerosols and cloud impacts deeply, and a few theories have emerged about the effects of aerosols on a cloud based on this past research. This study considers three theories: the direct, indirect, and semi-direct effect of aerosol forcing (figure 2.2).

Figure 2.1: Diagram showing thunderstorm evolution and electric field formation processes. (Dye et al. 1986)
2.2.1 The Direct Effect

The direct effect (DE), also known as aerosol radiative forcing, is centered on the thought that aerosols create direct radiative forcing. The DE is one of the more widely discussed aerosol effects (Sekiguchi et al. 2003, Papadimas et al. 2012, Podgorny and Ramanathan 2001, Benedetti and Vitart 2018) and assumes an impact on overall cloud formation processes. For the DE to occur, aerosols are directly altering the radiative properties of an area by absorbing sunlight and directly heating the atmosphere or an atmospheric layer while often promoting cooler temperatures at the surface (Koren et al. 2004). These forcings can occur from the atmosphere-surface, top of atmosphere, or atmospheric column prospective (Li et al. 2017). Direct radiative forcing from smoke aerosols varies between -0.2 and -1.1 Wm$^{-2}$ on average (Chylek and Wong 1995). This direct radiative
forcing, through scattering or absorption, can reduce incoming solar radiation. The reduction of solar radiation can suppress surface heating and alter production of convective eddies and upward motion (Figure 2.2, Wang et al. 2014).

2.2.2 The Indirect Effect

Also known as aerosol-cloud interactions (ACI), the indirect effect (IE) of aerosol forcing impacts clouds by directly changing the cloud’s microphysical processes. One sub-classification of the IE is the Twomey effect. The Twomey effect states that anthropogenic aerosols impact cloud albedo. As a cloud’s CCN increases, there are more smaller drops that form. These more smaller drops increase the ability for radiation to reflect and scatter when compared to a cloud of similar size with a lesser number of CCN (Twomey 1974, herein T74). If an average of 10% increase in nuclei occurred, a 2.5% increase in optical thickness would result with slower changes as the cloud thickness increases (T74). Cloud droplets coalescence and ice precipitation formation processes are inhibited in clouds of higher pollutant amounts (Rosenfeld 2000). In T74, analysis of air over an area was conducted prior to a wind shift as well as after the shift occurred. The clean air prior to the wind shift has much smaller nuclei concentrations when compared to the more urban-industrial air present after the wind shifted (Figure 2.3, Twomey 1974).

Another sub-effect is the Albrecht effect, also known as the cloud lifetime or second indirect effect. This effect considers marine stratocumulus and the formation of drizzle. In stratocumulus, a decrease in droplet size due to an increase
in CCN results in a liquid water content increase within the cloud. As a result, stratocumulus lifetimes are longer and come with an increase in albedo affecting Earth’s radiation budget (Albrecht 1989). The Albrecht effect is also believed to cause increased cloud heights as well due to the impact that the inclusion of more CCN has on large-scale precipitation processes (Figure 2.2, Pincus and Baker 1994).

Figure 2.3: Schematic showing the impact of polluted air on cloud nuclei concentration as a wind change due to sea breeze altered origins of air over the area. Very clean air was replaced by urban-industrial air from an area with coal mines, power stations, and other industries. A noticeable difference in nuclei concentration occurs with the presence of these air pollutants. (Twomey 1974)
2.2.3 The Semi-direct Effect

Lastly, the semi-direct effect (SE) is related to the aerosol effects which impact a cloud thermodynamically. The SE is thought to be driven primarily by decreases in low to mid-level clouds. These clouds help warm the system and alter the vertical aerosol atmospheric heating profile (Figure 2.2, Allen et al. 2019). As aerosols impact the vertical atmospheric heating profile, this in turn impacts stability and circulations. Aerosol coverage reduces cloud cover and produces heating at the top of the atmosphere, causing cloud burn-off. At the regional scale, aerosol solar absorption reduces daytime cloud coverage by nearly half (Ackerman et al. 2000).

When considering thunderstorms and how aerosols impact thunderstorms in particular, there are mixed results on the aerosol impact to lightning production. Some studies suggest that the greater CCN concentration tends to lead to greater lightning activity, although this relationship has a large sensitivity to the multiplication of ice (Mansell and Ziegler 2013, Yang and Li 2014, Liu et al. 2020), while others suggest that aerosol load reduces lightning activity (Yang et al. 2013, Tan et al. 2016, Altaratz et al. 2017). These aerosol impacts on clouds can produce differing results. For each effect, there are alterations to the overall storm-scale environment in some way. Changes in that storm-scale environment are greatly impacted by CCN availability and therefore impact the overall chances of charge separation and lightning formation. CCN differences in a cloud or warm
layers can alter storm characteristics which affect charge separation and lightning production.

2.3 Charge Separation and Lightning Formation

There are several mechanisms which impact charge separation processes in a cloud. Here, the processes discussed include the collisional charging mechanisms of selective ion capture, inductive charging, and non-inductive charging.

2.3.1 Selective Ion Capture

One of the earliest described suggestions for charge separation, selective ion capture, proposes that hydrometeors which are precipitating would become polarized as they fall given there is a positive charge present above ground during fair weather (Wilson 1929). This theory, typically called the Wilson selective ion capture mechanism, requires a suitable amount of electric attraction and fall speed for the hydrometeor to capture the single polarity ions (MacGorman and Rush 1998, herein MR98). The differing velocities of the hydrometeors and ions would create a region of negative charge along the bottom of the cloud. This mechanism is hypothesized to be unable to exist without the active presence of another charge separation mechanism. It is thought to only be impactful in storms that are weakly electrified as models estimate the electric field generated by this process alone would be at least an order of magnitude smaller than what is commonly present in a thunderstorm (Takahashi 1978, Chiu and Klett 1976).
2.3.2 Inductive Charging

The inductive charging mechanism is another method of charge separation. This theory assumes an ambient electric field and what occurs as hydrometeors fall through that field. As hydrometeors are falling, they come in contact with the polarized cloud vapor and assumedly become polarized as well (Sartor 1954). Charges are exchanged during this interaction as charge from the hydrometeor transfers over to the cloud vapor or a smaller ice crystal. If the cloud vapor rebounds, it keeps the charge it gained. When considering two hydrometeors, the process is further complicated as the interaction between the two hydrometeors must be considered as well as their interactions with cloud vapor or ice (Sartor and Helsdon 1981, Caranti et al. 1985). This mechanism is also thought to be a small part of the overall charging process as some investigations suggest thunderstorms cannot produce sufficient charge when considering this mechanism alone as well (MR98).

2.3.3 Non-inductive Charging

A last method of charge separation which is inferred to be important here is the non-inductive charging (NIC) method (Workman and Reynolds 1950, Reynolds et al. 1957, Illingworth 1985, Saunders 1993, Saunders and Peck 1998, Khain et al. 2001). This method considers any polarization which forms without the influence of an electric field, primarily considering a process commonly regarded as the non-inductive, graupel-ice collisional mechanism. This mechanism
and the NIC method overall is important here because the connection between updraft strength and lightning potential (Fuchs et al. 2018, Courtier et al. 2019, Deierling et al. 2008) is a result of the processes which contribute to this method of charging.

Workman and Reynolds (1950) suggested that the ice-water interface of freezing water was where charge separation occurred if the water originated from glazed ice that formed on the ice crystals. This was further supported to show that the graupel gains more charge when it is growing by riming and coming in contact with ice crystals, hence the graupel-ice collisional mechanism consideration (MR98). Dye et al. (1986, 1988) also discusses the charge separation mechanisms considering the region of cloud typically named the mixed-phase region: the region of the cloud where ice, graupel, and supercooled liquid water are able to coexist.

![Figure 2.4: Schematic showing inferred mass and charge transfer mechanisms between graupel and ice before, during, and after the charge transfer occurs. (Nelson and Baker 2003)](image)
This graupel-ice mechanism has been observed and modeled in past studies. It has been shown that more charge separation occurs when graupel undergo riming/colliding with ice crystals (Reynolds et al. 1957). Figure 2.4 shows this visually as Nelson and Baker (2003) describe how the mass and charge transfer occurs between the graupel and ice crystals. Others follow down the same path of thinking by finding that the non-inductive charge transfer occurs when there are ice crystals colliding with and separate from rime icing hailstones (Illingworth 1985) as well as showing that charging by hydrometeor collisions is adequate for thunderstorm electrification, with some assumptions (Saunders 1993).

The polarity of charges reverses given certain temperatures, liquid water content, and relative humidity amounts. Past studies have shown that, near the -10°C level, charges reverse under certain circumstances (Takahashi 1978, Saunders 1991, Pereyra et al. 2000) while others suggest reversal temperatures of around -21°C (Jayaratne et al. 1983, Keith and Saunders 1989). These studies all suggest that reversal temperature is varied with liquid water content (LWC), meaning LWC and temperature were thought to be the main factors in determining the key reversal temperature level. Later research, however, found that relative humidity was also influential on this overall process (Berdeklis and List 2001). Figure 2.5 helps to visualize how the overall reversal temperature process works. Above this level, larger, riming precipitation ice gains a net negative charge while smaller, non-riming ice crystals gain a net positive charge while below this temperature, larger, riming precipitation ice gains positive and smaller, non-riming ice crystals gain negative (Saunders 1993).
This is also related to accretion and the rime accretion rate (RAR) as a function of temperature (Saunders and Peck 1998, Brooks et al. 1997). The reversal temperature and RAR is instrumental in the formation of a tri-pole structure due to the charges and their typical distribution based on the outcomes of these processes (Simpson and Scrase 1937, Williams 1989, Saunders 1993). The general tri-pole structure has a lower positive charge region, main negative charge region, and upper positive charge region (Williams 1989, MR98, Mansell et al. 2010). Mansell et al. (2010) discuss this typical setup and a potential way this setup can evolve over time, seen here in figure 2.6. Here, the lower positive charge region formed as a result of graupel, main negative charge region formed
due to ice crystals and graupel, and upper negative charge region formed because of ice crystals and snow.

Figure 2.6: Diagram showing layer net charge and evolution through time as lightning occurs. (Mansell et al. 2010)

This layer formation is primarily due to differing terminal velocities of ice crystals, graupel particles, and snow as well as related to the temperature profile of the cloud. The larger graupel has a larger terminal velocity, typically on the order of 1.3 – 7.8 m/s, than the smaller ice or snow, typically 0.03 – 1.3 m/s (Sokol et al. 2020). These differing terminal velocities create congregation of these similarly-sized hydrometeors along the top and bottom of that portion of the cloud. The hydrometeors have exchanged mass and electrons during this time, creating these charged hydrometeors. As the terminal velocities limit the heights to which certain hydrometeors can go, which is also dependent on the strength of the updraft as well, the similarly-sized hydrometeors begin to collect within
certain regions of the cloud. As previously mentioned, the smaller hydrometeor typically gains a more positive charge while the larger hydrometeor typically has a more negative charge above the reversal temperature while below that temperature level, it is commonly the larger graupel which maintains the positive charge and ice is then negative. These hydrometeors separating within these cloud layers starts to form charged layers within these respective layers and subsequently the overall charging in the cloud. Lightning then occurs as these charging layers develop.

2.4 Charging and Wildfire Relationship

When cloud-to-ground lightning (CG) occurs, this is due to the presence of a strong electric field between charges of opposite sign. As a cloud forms these layers, the lowermost portion of the tri-pole charging region is typically smaller in magnitude. For the electric field to return to its normal state, the lightning must go somewhere. Thus, CG lightning occurs as a result. When lightning forms and comes in contact with the ground, LIW formation can then be considered. Given optimal ground conditions, some atmospheric parameters to consider for LIW creation include dewpoint depression, lapse rate, and the potential for dry lightning, with dry lightning being the parameter of most importance (Dowdy and Mills 2012). Since the area was so hot and dry after the recent heat wave with persistent high pressure, the surface conditions supported a lightning-initiated wildfire outbreak event. So, with the presence of dry lightning, potential for LIW exponentially increased.
Chapter 3. Data and Methods

For a comparison of these case days, a multi-sensor approach is necessary to allow for complete analysis of each topic area. Lightning data is necessary when considering the potential for wildfire, models are necessary to understand potential synoptic scale differences between the cases, single-cell analysis is necessary to understand storm-scale differences, an understanding of current surface conditions is necessary, sounding and Light Detection and Ranging (LiDAR) data is necessary for vertical analysis of the atmospheric column, and aerosol optical depth (AOD) and air quality data is necessary to diagnose potential aerosol differences. These data must be gathered in this manner to allow for full consideration of any vertical, horizontal, or temporal changes across the domain during the time of interest.

3.1 Atmospheric Forecast Models

Model data were considered to understand overall synoptic patterns for each of the case days. Here, the High-Resolution Rapid Refresh (HRRR v3, herein HRRR) operational model data were considered. HRRR data are available on an hourly basis and gathered from the University of Utah’s archive (Brian Blaylock, HRRR Webpage).
These data were selected to analyze synoptic height patterns present at the initialized model runs, which are on a 3 km grid with radar data assimilated every 15 minutes. The HRRR data was displayed using Python and Pygrib, a Python GRIB reader package, in the Plate Carrée cylindrical projection (Jeff Whitaker, PyGRIB Webpage).

Synoptic height patterns were assessed using the geopotential height field provided at hourly increments using HRRR data throughout the day. These data were plotted and viewed using the Python Matplotlib package (Hunter 2007).

3.2 Surface Observations and Atmospheric Profiles

Remote Automatic Weather Stations (RAWS) data were considered when assessing surface conditions. This dataset, with data retrieved from the Western Regional Climate Center’s (WRCC) archive, is a set of data collected mostly near areas with fire danger (WRCC, RAWS Webpage). These RAWS units provide many helpful observations, like providing access to air temperature, relative humidity, wind, precipitation, and fuels information, among others. These surface observations are very important when assessing the risk of wildfire with regards to available fuels at the surface as well as conditions which would allow for quicker fire spread, like increasing winds or no surface precipitation. RAWS sites, which covered the domain of interest for each day, were chosen to best represent the actual present conditions for each day. These sites were chosen to try and get a better understanding of the whole study domain and the sense of spatial variability across the domain.
Sounding data was downloaded from the University of Wyoming Department of Atmospheric Science’s upper air archive, which houses data from the NWS upper air sites (University of Wyoming, Upper Air Webpage). NWS soundings for the Oakland, California upper air site were visualized and analyzed using Python and the Python package MetPy (Unidata, MetPy Webpage). Soundings near the coastal region of California were modified to mitigate influences due to the presence of a coastal marine layer. To do this, these soundings were started at the closest pressure above the 850 mb level as this was near the height of the coastal marine layer over the area.

Air quality information for each of the case days is helpful to consider. This dataset provides information which can show potential differences at the surface level. Here, particulate matter is considered as the particulate matter quantity is a good representation of the surface smoke and aerosol presence. Daily PM2.5 data for each of the cases were downloaded from the United States Environmental Protection Agency’s (EPA) Outdoor Air Quality data archive (EPA, Air Quality Webpage). Here, the EPA’s data was downloaded and assessed using the Python Pandas package (The Pandas Development Team, Pandas Webpage) and Matplotlib (Hunter 2007).

3.3 Radar

Radar data for this study included the Next Generation Weather Radar (NEXRAD) data. This radar system, operating S-band Doppler weather radars which are commonly known as Weather Surveillance Radar, 1988, Doppler (WSR-
88D) radars, was used to view and interpret the level 2 reflectivity radar data. These data were accessed through NOAA’s NCEI data archive (NOAA/NCEI, NEXRAD Webpage). These data were visualized using Python and the Py-ART radar analysis package (Helmus and Collis 2016).

These data were used in conjunction with the Multi-Radar Multi-System (MRMS) radar data as well. MRMS is a National Severe Storms Laboratory (NSSL) product built on components from the National Mosaic and Multi-Sensor QPE (NMQ) (Zhang et al. 2011) and Warning Decision Support System-Integrated Information (WDSS-II) (Lakshmanan et al. 2007) systems. The MRMS severe weather products use WDSS-II while MRMS QPE products mostly consider the NMQ QPE. The MRMS system operates on data from multiple radars, observations, detection systems, and models allowing for a multitude of MRMS products to be generated. The MRMS gridded data is on a 1 km grid with 2-minute update cycles and returns data for 31 different vertical levels. These data were downloaded from Iowa State University’s Iowa Environmental Mesonet hourly archive (Iowa Environmental Mesonet, MRMS Webpage). The GRIB format for these files can be read by Python and the Python package PyGRIB (Jeff Whitaker, PyGRIB Webpage). There were several parameters used from this dataset including: composite reflectivity, vertically integrated ice, 18 dBZ echo top, 30 dBZ echo top, -10°C reflectivity, and quantitative precipitation estimate over the past hour. Composite reflectivity was chosen to use for cell tracking and total cell number. It must be noted that the composite reflectivity returns are maximum values for each grid point, which is not as useful for assessing certain vertical
layers of the atmosphere. This is also used as a comparison of maximum reflectivity values for each day. Vertically integrated ice (VII) is used to assess ice mass which is related to lightning potential (Mosier et al. 2011, Seroka et al. 2012, Bogel and Kredensor 2020). The 18 dBZ echo top is used as an estimate of actual echo top height (Lakshmanan et al. 2013) as it is related to the typical size and concentration of pristine ice crystals. The 30 dBZ echo top is used as it is typically a good estimate of graupel location in the cloud as it is the minimum reflectivity which indicates the presence of larger precipitation ice, such as graupel, with the necessary concentration to support electrification in the charging zone (Woodward et al. 2012). These heights are important because as the charge regions form in areas of higher heights, temperatures in those areas are cooler. Cooler temperatures are necessary for the creation of the ice crystals and graupel needed for charge separation.

The -10°C reflectivity heights were used to help understand layers most conducive to charge separation and lightning formation (Woodward et al. 2012). The -10°C reflectivity height is important because at this temperature level and cooler, frozen drops and graupel can be present in the cloud (Carey and Rutledge 2000). The past hour’s quantitative precipitation estimate is extracted to identify estimated surface precipitation based on radar data. This is an interesting parameter to use when considering the overall structure of the thunderstorm cells from the radar’s perspective. The model estimated freezing level information is also extracted to help in assessing mixed-phase depths given the heights of the 18 and 30 dBZ echo tops. For this study, mixed-phase depth is the volume of cloud
between the freezing level and estimated cloud top, commonly referenced to the 18 dBZ echo top height.

3.4 Satellite Data

The Visible Infrared Imaging Radiometer Suite (VIIRS) instrument on the Suomi-NPP satellite provides daily global AOD measurements which can be downloaded from the NOAA CLASS system (NOAA, VIIRS Webpage). This dataset provides daily observations, in the absence of unfavorable and cloudy circumstances, which are at 750 m resolution at nadir. When clouds are present, AOD cannot be retrieved. AOD is a measure of the extinction (scattering and absorption) of light by aerosols. AOD is a quantitative measurement related to the amount of aerosols in the atmosphere. As the AOD value increases, the extinction of light increases as well. This is a key reason why AOD is being examined here, as the lack or presence of the sun’s rays over an area can significantly alter the storm environment due to scattering and absorption of solar energy. These data were viewed with assistance from the NOAA/NESDIS Center for Satellite Applications and Research.

Cloud-Aerosol LIDAR Infrared Pathfinder Satellite Observations (CALIPSO) level 2 data were used to assess the vertical feature mask (VFM) information over the domain. The VFM, which identifies aerosol coverage in the vertical extent, helps to diagnose aerosol layers and the heights where aerosol layers are present. These data were downloaded in HDF format from the NASA Atmospheric Science Data Center’s EarthData archive (NASA Atmospheric Science Data Center, 24
CALIPSO Webpage) and viewed using Python, and the package PyHDF (Andre Gosselin, PyHDF Webpage), as well as an inspiration from NASA DEVELOP’s VOCAL package (NASA DEVELOP, VOCAL Webpage). In conjunction with an AOD, VFM helps to display differences in overall spatial distribution of aerosols which may impact storm formation based on the aerosol concentration and distribution within a certain atmospheric column.

3.5 Lightning Data

Vaisala’s National Lightning Detection Network (NLDN) data (Orville 2008, Cummins and Murphy 2009, herein CM09) were used to assess present CG lightning, overall totals of lightning, and the lightning’s estimated location. NLDN data is retrieved via electromagnetic wave from the lightning itself using the time-of-arrival system (Cummins et al. 1998). The NLDN system can pick out the CG lightning with a remarkable detection efficiency of 90-95% (CM09). Fields used in this study were day, time, latitude, longitude, polarity, and cloud-to-ground/in-cloud discriminator. Data were grouped into CG vs. non-CG and divided up for each cell based on the bounds and nearest time derived from cell tracking. Consideration was also given to polarity distributions as some studies suggest positive strikes may be more efficient at igniting wildfires (Nauslar 2014). The total cell counts and per minute counts were also discussed and were derived from a combination of the NLDN and single-cell data.
3.6 Study Area and Grid Formation

The location of convective cells for each day varied, so using data from one single radar site would not provide valuable information to compare convective cells. Because of this, the study region, shown in figure 3.1, for the first case is a 150 km by 150 km grid box at 1 km resolution centered on the San Francisco, California radar (KMUX) at 37.155, -121.898 while the second case has an identical grid centered on the Beale Air Force Base radar (KBBX) at 39.496, -121.632. All gridding and tracking methods for each day were held constant, with the only difference being in the radar change for each location. The grids were created using Py-ART and Py-ART’s grid_from_radar function. The grid had a shape of 300(x) by 300(y) by 19(z) with limits of -150,000, 150,000 in the x and y and z limits of 0, 19000. This grid draws inspiration from the MRMS data, as the MRMS comes with 19 z levels. These z levels are just levels which extend into the vertical, i.e. height coordinates. The grids are being given x and y radii of 150 km because this radius considers the WSR-88D’s area of best coverage. The radar’s best coverage area is to approximately 80 nautical miles, or approximately 148 km. Using 150 km as the radius allows for inclusion of this best coverage area.

The grid is at 1 km resolution. The radius of influence for the grid was the roi_func=`dist` method. This means that the radius of influence grows with the distance from each radar. Lastly, the algorithm used for gridding was gridding_algo=`map_gates_to_grid`. This maps each radar gate onto the grid using the radius of influence.
Figure 3.1: Study region showing the KMUX and KBBX radar location as well as the OAK upper air sounding site.

3.7 Storm Tracking

To compare the cases, storms were looked at on the cellular level to assess storm characteristics for each day. A cell was defined as any location where the composite reflectivity value was greater than or equal to 30 dBZ at or adjacent to the previously defined cell center grid at 4 km in altitude.
All original NEXRAD radar scans were converted to these previously described Py-ART grids. These grids were downloaded in the NetCDF format and read in by the TINT storm tracking algorithm (Raut et al. 2021). This algorithm is a storm cell tracking package led by the Data Informatics and Geophysical Retrievals (DIGR) Group at the Argonne National Laboratory. Most of the preset variables stayed the same, with just two major changes. The minimum size was changed to 20 square km based on the average thunderstorm size of around 24 km. This provides a small buffer to make sure smaller cells are counted due to the differing magnitudes of storm cells. The other change is the value for the height was changed to 4000 m. The aerosol coverage is mostly in the 3-5 km range across the study domain. This change to 4 km in the storm cell tracker allows for cells which were present in the aerosol-rich atmosphere to be the ones identified for the 24th case, while using the same height on the 16th case allows for a look at that day’s storm cells considering the clearer environment at the same vertical location. The tracker then interpolates and finds the closest radar height to that pre-defined height. All of the Py-ART grids are run through the tracker to identify cell centers for each radar scan. These data are then extracted and saved for future use.

Knowing the cell centers and times, bounds can now be inferred. To find the bounds, MRMS composite reflectivity was used. Composite reflectivity was used because it returns the maximum values of reflectivity for each grid point, which is helpful in identifying cell locations. Note that since these are maximum values, assessing reflectivity distributions or using reflectivity in conjunction with
another dataset is best done with another reflectivity product which preserves the vertical variations in reflectivity. A 2 grid box buffer was added around the cell to ensure sufficient inclusion of cell values. These cell bounds were then saved for the extraction of MRMS values as well as other values for individual cell analysis.
Chapter 4. Results

To describe these cases in a quantitative sense and test the hypotheses, gathered data are described at varying scales and height ranges to assess overall conditions. For the differences in mixed-phase storm-scale characteristics, single cells were analyzed. The time of the very first tracker-identified cell was 0105 UTC on August 16th and 0210 UTC on August 24th. The last identified cell time was 2106 UTC and 2156 UTC, respectively. To assess the presence of an enhanced concentration of aerosols, aerosol data was analyzed to identify potential differences in case days. All of these data are considered to try and understand the multiple influences which might have had an impact on the differing case outcomes.

4.1 Land Conditions

During the late summer months of 2020, persistent high pressure, record-setting maximum temperatures, and numerous heat waves occurred across California. These record-breaking events helped influence the Climate Prediction Center (CPC) as they estimated persistent drought conditions would be present throughout the month of August 2020, shown in figure 4.1, with the likelihood of
more drought development throughout the month of August over small portions of central California (National Weather Service Climate Prediction Center).

![U.S. Monthly Drought Outlook](image)

**Figure 4.1:** U.S. monthly drought outlook for August 2020.

Due to the present hot and dry conditions, land conditions stayed relatively the same throughout the whole month of August as can be seen in the RAWS data. One of these sites in the study area is the San Jose site. The San Jose, California RAWS site returned fuel moisture values of 9.5% fuel moisture average on August 16th. For the 24th, the RAWS fuel moisture was 13% average. Data for this site can be seen in figure 4.2. Although a minimal 3.5% increase in fuel moisture occurred from the first case day to second case day, these values are still much below the 30% threshold which assumes dead fuels.
Figure 4.2: Fuel moisture time series for the San Jose, California RAWS site. August 16th reported maximum fuel moisture average of 9.5 percent and August 24th reported maximum fuel moisture average of 13 percent. Any fuel moisture below 30% is essentially considered dead fuels.

Because fuel moisture patterns consistently reported less than the 30% threshold and there were small amounts of rainfall over an already drought-stricken land, land conditions were not considered. There were only eight days between the cases, so the land surface did not rapidly change in this short amount of time with the given weather patterns over the area.

4.2 Synoptic Pattern

The environment was examined to see if a differing overall synoptic pattern or changes in the thermodynamics of the area have resulted in different storm formation.

Recent research has provided a model setup for favorable conditions needed for successful dry lightning and, therefore, dry lightning fire ignition (Kalashnikov
et al. 2022, herein K22). This paper discusses this case in particular, as well as describing this setup as a cluster 1 variety. This variety is described as having increased dry lightning likelihood over the domain with the largest median spatial extent and lowest median elevation of dry lightning amongst all of the different clusters discussed in the study. K22 describes this setup exactly, with ridging over the western portions of the continental United States and troughing offshore. This transitional pattern is most typically associated with widespread dry lightning outbreaks. This overall synoptic pattern is conducive to dry lightning based on past dry lightning events.

4.2.1 500 mb

The 0000 UTC model runs in figure 4.3a and 4.3b show height patterns over the area. August 16th has a near north-south vertical pattern with heights increasing from 5900 to 6000 gpm in an eastward fashion. The 24th has a c shaped height pattern over the area with the higher heights in the central eastern portion of the area with heights near 5940 gpm compared to the contrasting 5840 gpm regions on the northwest and southwest corners of the region.

By 0200 UTC, areas to the west of the region on August 16th in figure 4.4a have decreased 30 gpm as areas to the east of the region have remained constant. On August 24th at 0200 UTC in figure 4.4d, the northwestern lowered heights of 5840 gpm are progressing towards the east while the more southern region is holding near constant. The highest heights across the region of 5940 gpm are collecting while preparing to exit the region. By 0600 UTC, heights have
gradually increased to a range of 5910 to 5970 gpm on the 16th figure 4.4b and 5860 to 5910 gpm on the 24th figure 4.4e as slight movement of higher heights towards the east is occurring as the trough progresses over the area. This small "kink" in the wave pattern signals an area of increased upper-level disturbances and, therefore, increased potential for thunderstorm formation. By 0900 UTC in figure 4.4c and 4.4f, not many changes occur with both patterns shifting slightly towards the east.

By 1100 UTC, a small "kink" has also formed on the 16th figure 4.5a, just south of the 24th figure 4.5d location. These differing potential shortwave locations align well with storm cell location as the cells on the 24th had a more northerly origin when compared to the more southerly placement of the 16th shortwave. By 1300 UTC, however, a small break in the wave occurred near the Bay area on figure 4.5b as a smaller break occurs in a similar location on the 24th figure 4.5e. At this time, heights on the 24th are slightly decreasing from
Figure 4.4: Geopotential height patterns for 0200, 0600, and 0900 UTC on August 16 and August 24.
5900 at 1100 UTC to 5870 gpm at 1300 UTC as the trough prepares to exit the area. Not much change occurred by the 1500 UTC hour on figure 4.5c and 4.5f as the break in the wave makes a more northerly assent and makes it fully onto land.

By 1700 UTC, both days in figure 4.6a and 4.6d start to form a north-to-south oriented height pattern with increasing heights towards and moving east with 5895 gpm at the far western portion of the region to 5890 at the far eastern portions on the 16th while the 24th increased from 5810 along the western side to 5900 in the bottom eastern corner.
Heights on each day then began to increase over the southern portions of the region as the 19 UTC hour approaches in figure 4.6b and 4.6e. By 22 UTC in figure 4.6c and 4.6f, the California regions south of Nevada were under the highest areas of heights for each day of around 5985 gpm and just south of the compact ridging present from the northwest corner of the area stretching down towards the southwestern corner of Nevada. This compact ridging is shown for both days and is just south of this area on the 24th with values of around 5900 gpm present.

By the end of the day in figure 4.8a and 4.8b, not much has changed as both days have height patterns showing maximums in the southeastern corner,
Figure 4.7: Geopotential height patterns for 2300 on August 16 and August 24.

5990 for the 16th and 5900 for the 24th, with minimums in the northwestern corner of 5895 for the 16th and 5820 for the 24th. Slight differences of 5930 on the 16th to 5840 on the 24th, however, are present offshore. The maximum height difference for each day is 62 gpm difference as the max over the area was 5979 gpm for the 16th and 5917 gpm for the 24th.

4.2.2 850 mb

At the 850 mb level, by 4 UTC in figure 4.8a and 4.8c, an area of higher heights of around 1560 gpm is to the south of Nevada. There is also a maximum in this similar location on the 24th as well of around 1520 gpm. By 6 UTC on the 16th in figure 4.8b, a small maximum of 1550 gpm has developed over the oceanic regions of the domain as well as across the northeastern quadrant of 1560 gpm. For the 24th by this time in figure 4.8d, there is still that maximum height cluster south of Nevada around 1530 gpm and in the more interior portions of California at 1510 gpm.
Figure 4.8: Geopotential height patterns for 0400 and 0600 UTC on August 16 and August 24.
By the 8 UTC timeframe in figure 4.9a and 4.9c, the 16th maximums have decreased their spatial coverage while the 24th has increased its maximums to 1545 gpm and slightly increased maximum spatial coverage as well. By 10 UTC in figure 4.9b and 4.9d, there was a horizontally-oriented area of max heights around 1550 gpm near the Monterey Bay area. This corresponds not only with the time of most intense storms moving onshore but also is near the estimated fire start time. The 24th had developed an area of more concentrated higher heights of 1540 gpm just west of the Nevada/California border. This also corresponds to the area where the cells were most prone to initiate on the 24th.

By 12 UTC on the 24th in figure 4.10b, this max has slightly dissipated to 1525 gpm and broadened to cover most of the northeast quadrant of the region. The 16th in figure 4.10a has an area of maximums near 1540 gpm that has moved northward slightly to the San Francisco Bay area. By 14 UTC in figure 4.10c and 4.10d, both days see the eastern portion of the region starting to see increasing heights as the max regions begin to dissipate completely and blend in with the larger scale patterns over the region. By 16 UTC in figure 4.10e and 4.10f, a more distinct maximum again occurs in the northerly portion of the region, with a more distinct max on the 16th of 1565 gpm than the 24th of 1540 gpm as well as a more easterly placement for the 24th max. By 19 UTC on the 16th in figure 4.10g and 4.10h, another max of 1580 gpm has occurred in the central California area just north of Sacramento. For the 24th, higher max areas of 1530 gpm are present consistently across the interior portions of California.
Figure 4.9: Geopotential height patterns for 0800 and 1000 UTC on August 16 and August 24.
Figure 4.10: Geopotential height patterns for 1200, 1400, 1600, and 1900 UTC on August 16 and August 24.
By 21 UTC, the 16th in figure 4.11a has seen some increase in areas south of the bullseye as coastal regions are now having higher heights as well. At this time on the 24th in figure 4.11b, areas are beginning to decrease in height as smaller maximum regions begin to develop to the south and north of the interior California regions. By the end of the day for each day in figure 4.11c and 4.11d, higher heights of 1560 for the 16th and 1520 for the 24th exist for both of the days over the interior California regions of the study area and near the California/Nevada border, where a strong gradient occurs for both days. The maximum height difference for each day is 56 gpm difference as the max over the area was 1585 gpm for the 16th and 1529 gpm for the 24th.

### 4.2.3 1000 mb

For the start of the day in figure 4.12a and 4.12b, lower heights were in the interior of California for both days with the main differences occurring over the very northeastern portion of the domain. There is a small max of 110 gpm over the corner of the Nevada/California border and an area of minimums of 80 gpm just east of that for the 24th. By 2 UTC in figure 4.12c and 4.12d, the 24th minimums have shifted towards the west as well as the max showing on the 16th elongating along the eastern mountainous regions along the east as well.

By 6 UTC in figure 4.13a and 4.13d, lowered heights all along the central California region as low as 80 gpm have formed as lowered heights of 70 gpm spread across California on the 24th as well. By 10 UTC in figure 4.13b and 4.13e, lowered heights of 80 gpm persist along the central California region on the
Figure 4.11: Geopotential height patterns for 2100 and 2300 UTC on August 16 and August 24.
Figure 4.12: Geopotential height patterns for 0000 and 0200 UTC on August 16 and August 24.
16th as well as all along the northern and northeastern portions of the area around 70 gpm on the 24th. By 12 UTC in figure 4.13c on the 16th, the heights have lowered even more, especially in the southern portions of the region to around 70 gpm. The 24th in figure 4.13f, conversely, has increased slightly to 65 gpm.

By 15 UTC in figure 4.14a and 4.14d, small maximums have developed along the northeast portion of the region on both days and small areas of minimums of 130 gpm on the 16th and 95 gpm for the 24th are present. By 17 UTC in figure 4.14b and 4.14e, regions of higher heights are still in the northern half of the domain near 135 gpm on the 16th and 95 gpm on the 24th with some lower values to the south, although these have started to increase a bit. By 20 UTC in
Figure 4.14: Geopotential height patterns for 1500, 1700, and 2000 UTC on August 16 and August 24.

Figure 4.14c, a distinct maximum of higher heights near 160 gpm is present on the 16th while there are lower heights to the south again. For the 24th in figure 4.14f, a smaller magnitude max of 95 gpm exists in the same general region as lower heights are to the south for this day as well.

By the end of the day in figure 4.15a and 4.15b, lower heights begin to creep northward with a strong gradient in height for both days just east of the Nevada/California border. A smaller region of minimums is to the south as well. The maximum height difference for each day is 39.5 gpm difference as the max over the area was 156.4 gpm for the 16th and 116.9 gpm for the 24th.
Using the random.sample provided in Python, and choosing a sample size of n=30, running the Mann-Whitney U test considering values for 1000, 850, and 500 mb throughout the whole day, the p-values returned are 0.371, 0.158, and 0.217 respectively. Given their values above 0.05, no statistical significance is given to the synoptic scale parameters. It must be noted, however, that this is a limited test and is testing the magnitude of the synoptic variables alone, and not their spatial variations.

These patterns are important here as the overall patterns which were present showed similar trends. These patterns matter to the study because, if vast differences were present, the synoptic setup and the variations which were present could have been a major, or even the only, factor contributing to the overall outcome. The synoptic pattern does not provide any drastic changes which might have signaled differences in synoptic pattern as a single cause in the case outcomes, however.
4.3 Thermodynamic Influences

So, given the outcome of the land and synoptic analysis, would thermodynamic differences have impacted the mitigation of lightning present on August 24?

When looking at the visual skew-T here in figure 4.16a, the August 16 environmental sounding is displayed while figure 4.16b shows the August 24 sounding. These soundings show the estimated parcel line given the surface estimated dewpoint and temperature as well as those estimated dewpoint and temperature lines from the balloon launches. For each of these soundings and their associated profiles, the estimated LCL, or estimate for cloud base, is 1.9 km on the 16th and 2.6 km on the 24th. For the level of free convection (LFC), or the level where the parcel begins to freely accelerate upwards until something hinders upward motion, this level is estimated to be approximately 2.8 km and 3.2 km on the 16th and 24th, respectively.

For the lower levels, the 0 km level in particular, the dewpoint depression, or the difference between temperature and dewpoint, is 38° difference (74°F temperature, 36°F dewpoint) on the 24th compared to 27° difference (76°F temperature, 49°F dewpoint) on the 16th. This increased dewpoint depression, or increased distance between dewpoint line and temperature line, on the chart signals drier lower-level air for the 24th case. A deep dry layer is present in the lower portions of the troposphere between the surface to approximately 2 km on both days. The August 24 event had a larger dewpoint depression, however, as there
Figure 4.16: Representative sounding from the Oakland, CA upper air site for August 16 and 24, 2020 at 12 UTC. The 12 UTC sounding was chosen due to the times when cells were present on each day. The green line identifies dewpoint, red line identifies environmental temperature, and white dotted line represents the parcel line when lifted from the surface.

was a median dewpoint depression of $12^\circ$ on August 24 compared to $8^\circ$ on August 16. This increased dewpoint depression displays increased potential for dry air evaporating precipitation throughout this lower tropospheric layer. This air can also be contributing to dry air entrainment causing more evaporation within the cloud or from the boundary layer at or near the cloud base. Evaporation is a cooling process which helps stabilize the atmospheric column.

There is a large difference between the days in the 3 - 6 km layer. The 24th has a layer of drying with a dewpoint depression of approximately $6^\circ$ difference at the lower portion of this layer around 4 km while the 16th has a much closer dewpoint and temperature line relationship with about a $1^\circ$ difference, signaling a higher relative humidity for the 16th in this layer. This continued difference
in dewpoint depression suggests additional opportunities for increased dry air entrainment for the 24th case along the cloud edge boundaries. Just atop these decreased dewpoints, the 24th then warmed and increased in dewpoint relative to the parcel line to bring both of these values near -10°C at approximately the 5 km layer, signaling a temperature inversion in this 5-6 km level of the atmosphere for the 24th. The 16th does not have this present, however. This inversion, and associated capping, helps to hinder convective growth and deep convective formation.

When considering convective available potential energy (CAPE) and its relationship to the increasing convective nature of storm cells, this inversion hinders healthy storm cell convective growth. The lower tropospheric deep dry layer and increased dewpoint depression also reduce CAPE and increase CIN by altering the parcel paths. Since the 24th event had that increased dewpoint depression, the effects were more pronounced on that atmospheric profile when compared to the 16th profile, however. Since CAPE is schematically defined as the area between the environmental temperature line and the assumed parcel path shown in the white dotted line, the inversion in that temperature line effectively removes a large portion of the area which could be used for CAPE area had the inversion not been present. When considering the convective inhibition (CIN) potential, CIN is a numerical representation of the amount of capping present. Capping describes the portion of the atmosphere that does not support the formation of lifting or deep convection. Calculated CAPE values for these soundings were also assessed. The most unstable (MU) CAPE value is the largest calculated CAPE value given
many varying parcel paths and CAPE calculations. MU CAPE values were 1287 J/kg on the 16th and 570 J/kg on the 24th while MUCIN values were -39 J/kg on the 16th and -4 J/kg on the 24th. Surface-based (SB) CAPE is the CAPE value calculated when raising a parcel from the surface. SB CAPE values were 779 J/kg and 521 J/kg on the 16th and 24th respectively while SB CIN values were -266 J/kg on the 16th and -463 J/kg on the 24th.

These values correspond with what is shown on the skew-T schematics as well. Larger portions of the area under the parcel line can positively contribute to the CAPE calculation on the 16th and smaller portions of this same profile contribute to CIN. There were notable differences in dewpoint depression in the upper levels as well, particularly in the 6 - 10 km range where increased temperatures and decreased dewpoints were present.

All of the thermodynamic influences which have been considered are summarized in table 4.1. CAPE and CIN were chosen because of their ability to diagnose the convective nature of the atmosphere. LCL was chosen to directly compare cloud structure. Shear and lapse rates were chosen to provide a better understanding of vertical thermodynamic trends. The parameters are assessed at the times available from the NWS regularly launched upper air sounding site of Oakland, California.

August 16 SB CAPE values were 176 J/kg and 258 J/kg greater for the 0 UTC and 12 UTC soundings for the days of interest while SB CIN was -774 J/kg and -197 J/kg smaller at these same times on the 24th. The MU CAPE was also greater on the 16th by 173 J/kg for the 0 UTC and 717 J/kg for the 12 UTC
sounding. The MU CIN was -189 J/kg smaller on the 16th at 0 UTC compared to -35 J/kg smaller at 12 UTC on the 16th. The LCL height is estimated to be 90 m higher on the 16th at the 0 UTC sounding while the 12 UTC sounding shows an LCL of 738 m higher on the 24th by this time. The surface to 6 km shear shows that at 0 UTC, the 16th had stronger 0-6 km shear by 9 kts while at the 12 UTC sounding the 24th had larger 0-6 km shear by 7 kts, reversing the previous pattern. The surface - 3 km lapse rate was 1.9 °C/km greater on the 16th at the 0 UTC sounding while this lapse rate was 0.5 °C/km greater on the 24th by the 12 UTC sounding. For the 3 - 6 km lapse rate, the difference was nearly identical with 0.2 °C/km greater on the 16th for the 0 UTC timeframe while there was a 0.3 °C/km difference by the 12 UTC sounding.

When considering these CAPE values, present values display a larger difference between the days at the 12 UTC sounding with larger magnitudes. This sounding also had potential for smaller impacts of CIN as well as a smaller difference between the CIN values for each day as well. This is present in both the CAPE calculations discussed here as well as overall CIN patterns considering the visual skew-T. The LCL height decreases throughout the day for both days, although it decreases a shorter distance for the 24th event. LCL eventually reaches heights lower than those at the 0 UTC sounding but is still well above the LCL for the 16th event. Shear decreases from the 0 to 12 UTC sounding for the 16th event were present while the 24th event had an increase in shear. The lapse rates decreased by 0.1 °C/km for the 16th event while remaining the same for the 24th event from 0 to 12 UTC.
This low-level dewpoint depression difference alters the parcel path that would form given the differing location of dewpoint and temperature lines present in the low-levels. Then, there would also be a difference in the associated adiabats which would be used to calculate the parcel path. This different parcel path, different lapse rate, and presence of an inversion would all impact the buoyancy and CAPE.

Dry air entrainment also occurs but isn’t accounted for in typical parcel theory as parcel theory commonly ignores this process. Because of parcel theory’s treatment of dry air entrainment, actual updraft strength would be less than the assumption from CAPE and parcel theory. This difference in actual and estimated updraft strength would be greater for the 24th case, i.e. suggesting an even weaker updraft than expected based on sounding values, because of the increased dewpoint depression and dry air present. Since both the parcel path and the lapse rates affect the buoyancy profile and CAPE, less CAPE and less buoyant air on the 24th, combined with increased dry air entrainment potential, would result in less deep, less vigorous convection on the 24th.
Table 4.1: Thermodynamic parameters assessed at the regularly launched NWS sounding times of 0 and 12 Z for August 16th and 24th as well as the 0 Z launch for the following day (17th/25th) for Oakland, California.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>0 Z</th>
<th>12 Z</th>
<th>0 Z</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>16 l 24</td>
<td>16 l 24</td>
<td>17 l 25</td>
</tr>
<tr>
<td>SB CAPE (J/kg)</td>
<td>306 l 130</td>
<td>779 l 521</td>
<td>532 l 90</td>
</tr>
<tr>
<td>SB CIN (J/kg)</td>
<td>-359 l -1133</td>
<td>-266 l -463</td>
<td>-256 l -312</td>
</tr>
<tr>
<td>MU CAPE (J/kg)</td>
<td>347 l 174</td>
<td>1287 l 570</td>
<td>1103 l 386</td>
</tr>
<tr>
<td>MU CIN (J/kg)</td>
<td>-209 l -20</td>
<td>-39 l -4</td>
<td>-10 l -131</td>
</tr>
<tr>
<td>LCL (m)</td>
<td>2992 l 2909</td>
<td>1892 l 2630</td>
<td>2132 l 1273</td>
</tr>
<tr>
<td>surface - 6 km shear (kt)</td>
<td>25 l 16</td>
<td>18 l 25</td>
<td>36 l 40</td>
</tr>
</tbody>
</table>

Continued on next page
Given these atmospheric profiles present with the drier surface layer on the 24th, a LIW outbreak seems more likely on this day based on the presence of increased dewpoint depressions. There is certainly potential, however, on both days for LIW given the development of the near-tropospheric dry layer which occurred for each. The largest difference in the two days is in the overall lightning potential, with the lightning potential being much higher on the 16th than the 24th.

There were notable differences in thermodynamic setups. This is important as thermodynamic differences such as these would support the creation of different vertical cloud structures and therefore support different lightning potentials for each day.
4.4 Thunderstorm Investigation

Each cell was investigated to assess its unique characteristics given the differences in thermodynamic profiles. For the purposes of this study, each time a cell was identified and given bounds, this single time and its data is called a cell snapshot. These are preserved to get a better understanding of the whole thunderstorm environment on each day. These data are summarized in the MRMS Parameters table 4.2. Considering all cells, August 16th had cells of higher cloud top heights, higher -10°C reflectivity, higher composite reflectivity magnitudes, and larger percentages of VII when compared to those cells identified on the 24th.

For the lightning data, each cell snapshot was investigated as well as whole cell characteristics on a per minute basis. Figure 4.17 shows cell CG lightning counts as a function of cell VII ice mass. There are stark differences in each day with the 16th event having much higher ice mass near 3500 million kg of ice and the 24th having around 250 million kg as well as CG lightning of 13 CG/minute/cell compared to 4 CG/minute/cell for the 24th. This figure also shows these same values in a zoomed in way to show the true variability in the two.

Figure 4.18 shows a frequency distribution for the number of negative polarity CG lightning flashes identified in each cell and the total cell counts for each cell. Again, the 16th values are dominating the plot and analysis with the
**Table 4.2:** MRMS parameters for the days of interest. These parameters include the data for the individual cell snapshots to preserve details throughout the whole lifetime of each cell.

<table>
<thead>
<tr>
<th>parameter</th>
<th>16th</th>
<th>24th</th>
</tr>
</thead>
<tbody>
<tr>
<td>number of total cells</td>
<td>527</td>
<td>160</td>
</tr>
<tr>
<td>number of cell snapshots</td>
<td>2218</td>
<td>699</td>
</tr>
<tr>
<td>reflectivity below 30 dBz max</td>
<td>0.00%</td>
<td>0.29%</td>
</tr>
<tr>
<td>reflectivity between 30-40 dBz max</td>
<td>10.55%</td>
<td>20.46%</td>
</tr>
<tr>
<td>reflectivity between 40-50 dBz max</td>
<td>54.82%</td>
<td>72.53%</td>
</tr>
<tr>
<td>reflectivity between 50-60 dBz max</td>
<td>32.87%</td>
<td>6.72%</td>
</tr>
<tr>
<td>reflectivity greater than 60 dBz max</td>
<td>1.76%</td>
<td>0.00%</td>
</tr>
<tr>
<td>VII present</td>
<td>83.54%</td>
<td>59.08%</td>
</tr>
<tr>
<td>average 18 dBz echo top</td>
<td>12.21 km MSL</td>
<td>10.52 km MSL</td>
</tr>
<tr>
<td>average 30 dBz echo top</td>
<td>8.94 km MSL</td>
<td>6.75 km MSL</td>
</tr>
<tr>
<td>average freezing level height</td>
<td>4.79 km MSL</td>
<td>4.70 km MSL</td>
</tr>
<tr>
<td>average -10°C reflectivity</td>
<td>35.99 dBz</td>
<td>31.58 dBz</td>
</tr>
</tbody>
</table>
Figure 4.17: Cell cloud-to-ground NLDN counts per cell per minute compared to overall cell mass for each day (a) and a zoomed in view of the same (b) displaying greater variabilities.

24th values not reaching above 2 negative CG flashes for any cell compared to values near 50 for the 16th.

CG lightning flash rate information at minimum, 25th percentile, median, 75th percentile, and maximum were retrieved for each dataset. These data and their returns can be viewed in figure 4.19 below along with the MRMS parameters and their percentile distributions. Many of the data values display very different trends for each day. VII has a 25th percentile of 2 for the 16th compared to 0 for the 24th. By the 75th percentile, August 16 returned a VII value of 64.875 million kg of ice compared to the 24th’s mere 7.5 million kg of ice. The VII, echo top, and lightning information seem to be the most varying of all these data.

Figure 4.20 analyzes the model estimated freezing level, 30 dBZ echo top height, maximum composite reflectivity, and -10°C reflectivity and the relationship these parameters have to ice mass.
Figure 4.18: The number of negative (a, b zoomed) and positive (c, d zoomed) CG lightning detected within each cell each day.
Figure 4.19: The percentile statistical analysis for each of the key MRMS variables. These include VII ($10^6$ kg), freezing level (km), -10°C reflectivity (dBZ), composite reflectivity maximum (dBZ), 30 dBZ echo top (km), 18 dBZ echo top (km), precipitation over the past hour (mm), and CG lightning (# of CG).

For the 30 dBZ echo top height / ice mass comparison, cells on the 16th were able to get to much higher echo top heights, some as high as 16 km, with much larger ice mass, nearly 1000 million kgs of ice. The 24th, in contrast, had 30 dBZ echo top heights reaching near 12 km with as much as 100 million kgs of ice. The ice mass is an order of magnitude different with upwards of 4 km difference in the echo top heights. The 30 dBZ is helpful here because it is an estimate of the most convective region of the cloud where graupel can exist given its typical concentration and size.

When considering maximum composite reflectivity, cells on the 16th were able to get to near 70 dBZ returns in some instances while the 24th had maximums of around 55 dBZ. The 16th and 24th had similar reflectivity distributions, the 24th just much smaller ice mass when compared to the 16th. Maximum composite

<table>
<thead>
<tr>
<th>parameter</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
</tr>
</thead>
<tbody>
<tr>
<td>VII (10^6 kg)</td>
<td>0</td>
<td>0</td>
<td>2</td>
<td>0</td>
<td>2</td>
<td>0</td>
<td>3.5</td>
<td>1.3</td>
<td>64.875</td>
<td>7.5</td>
<td>970.3</td>
<td>72.2</td>
</tr>
<tr>
<td>-10°C reflectivity (dBZ)</td>
<td>16th</td>
<td>24th</td>
<td>17.5</td>
<td>12</td>
<td>31.5</td>
<td>28.5</td>
<td>35.5</td>
<td>31.5</td>
<td>52.5</td>
<td>45.5</td>
<td>52.5</td>
<td>45.5</td>
</tr>
<tr>
<td>Composite reflectivity maximum (dBZ)</td>
<td>16th</td>
<td>24th</td>
<td>32.2</td>
<td>18</td>
<td>43.2</td>
<td>40.8</td>
<td>47.3</td>
<td>43.3</td>
<td>51.8</td>
<td>46.2</td>
<td>51.8</td>
<td>46.2</td>
</tr>
<tr>
<td>30 dBZ echo top (km)</td>
<td>16th</td>
<td>24th</td>
<td>1.25</td>
<td>1.688</td>
<td>7</td>
<td>5.333</td>
<td>8.333</td>
<td>6.5</td>
<td>11</td>
<td>11</td>
<td>11</td>
<td>11</td>
</tr>
<tr>
<td>18 dBZ echo top (km)</td>
<td>16th</td>
<td>24th</td>
<td>5.625</td>
<td>5.457</td>
<td>11</td>
<td>11</td>
<td>12</td>
<td>11</td>
<td>13</td>
<td>14</td>
<td>13</td>
<td>14</td>
</tr>
<tr>
<td>Precip over past hour (mm)</td>
<td>16th</td>
<td>24th</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>CG lightning (# of CG)</td>
<td>16th</td>
<td>24th</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>parameter</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
<th>16th</th>
<th>24th</th>
</tr>
</thead>
<tbody>
<tr>
<td>VII (10^6 kg)</td>
<td>0</td>
<td>0</td>
<td>2</td>
<td>0</td>
<td>2</td>
<td>0</td>
<td>13.5</td>
<td>1.3</td>
<td>64.875</td>
<td>7.5</td>
<td>970.3</td>
<td>72.2</td>
</tr>
<tr>
<td>-10°C reflectivity (dBZ)</td>
<td>16th</td>
<td>24th</td>
<td>17.5</td>
<td>12</td>
<td>31.5</td>
<td>28.5</td>
<td>35.5</td>
<td>31.5</td>
<td>52.5</td>
<td>45.5</td>
<td>52.5</td>
<td>45.5</td>
</tr>
<tr>
<td>Composite reflectivity maximum (dBZ)</td>
<td>16th</td>
<td>24th</td>
<td>32.2</td>
<td>18</td>
<td>43.2</td>
<td>40.8</td>
<td>47.3</td>
<td>43.3</td>
<td>51.8</td>
<td>46.2</td>
<td>51.8</td>
<td>46.2</td>
</tr>
<tr>
<td>30 dBZ echo top (km)</td>
<td>16th</td>
<td>24th</td>
<td>1.25</td>
<td>1.688</td>
<td>7</td>
<td>5.333</td>
<td>8.333</td>
<td>6.5</td>
<td>11</td>
<td>11</td>
<td>11</td>
<td>11</td>
</tr>
<tr>
<td>18 dBZ echo top (km)</td>
<td>16th</td>
<td>24th</td>
<td>5.625</td>
<td>5.457</td>
<td>11</td>
<td>11</td>
<td>12</td>
<td>11</td>
<td>13</td>
<td>14</td>
<td>13</td>
<td>14</td>
</tr>
<tr>
<td>Precip over past hour (mm)</td>
<td>16th</td>
<td>24th</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>CG lightning (# of CG)</td>
<td>16th</td>
<td>24th</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
<td>0</td>
</tr>
</tbody>
</table>
Figure 4.20: The maximum 30 dBZ echo top (a), composite reflectivity maximum (b), model estimated freezing level (c), and maximum -10°C reflectivity (d) per cell as a function of ice mass.
reflectivity is useful here because it gives some reference to the potential number concentration and size information of the cloud hydrometeors within the cell.

For the model estimated freezing level, the range of values is not that different. There is only 0.1 km height difference between the two. The freezing level heights are much more similar than the other variables. Freezing level is helpful to know when assessing mixed-phase depths as the freezing level is the assumed bottom of the mixed-phase region.

Lastly, for the -10°C reflectivity, values on the 16th have a near 55 dBZ maximum while on the 24th they are around 45 dBZ for a max. The higher the -10°C reflectivity is, the greater the potential for lightning as there is also potential for larger precipitation ice near this temperature level. These maximums show that, once again, there is a difference in the fundamental cloud properties of the upper-levels, particularly a difference in the levels of interest when considering the potential for lightning formation.

For the cellular characteristics, figure 4.21 includes correlation matrices for each day. The most correlated variables on the 16th are the 30 dBZ and -10°C reflectivity information. The least correlated variable seems to be the freezing level. For the 24th, the most correlated variables seem to again be the 30 dBZ and -10°C reflectivity information. This time, however, the amount of variables which are less correlated is much greater, especially for the precipitation estimate. These variable differences are one way to see how the cloud structure and the associated parameter returns were altered on the different days. Freezing level was one of the parameters which stood out on both charts. This value is a key
Figure 4.21: The correlation matrices for the MRMS variables for August 16th (a) and August 24th (b).

One as it was very similar on average for both days. The relationship it has with the other parameters changes greatly for each day, however. This is due to the decreased mixed-phase depth, decreased -10°C reflectivity, and decreased overall ice mass within the storm cells as a whole on August 24th.

Given the vast difference in thunderstorm quantity, lightning production, and microphysical parameters, some phenomena has to be occurring that can not
only impact but completely alter current atmospheric conditions, particularly in the upper levels where the greatest change between days is occurring.

4.5 Aerosol Contribution

When looking at the days through satellite data visually in figure 4.22, one can see that there is a noticeable difference in clarity with August 16 being generally clear while August 24 is very smoky. This can be seen in both the VIIRS true color satellite imagery as well as the local outdoor viewing webcam.

Although there is a noticeable visual difference, a smoky picture alone is not a way to actually quantify the air or the aerosol concentration in the air.
Figure 4.23: A visual depiction of aerosol optical depth for August 16 (A) and August 24 (B). AOD is a unitless measurement that quantitatively estimates the amount of aerosol based on the extinction of a ray of light passing through the atmosphere. As the extinction increases, higher aerosol is assumed in the area. Higher aerosol assumptions return a higher AOD value when the scattering is the highest.

4.5.1 Optical Depth and SpatialExtent

AOD values were assessed to better understand smoke concentration locations across the area. AOD plots presented here show the aerosol coverage and density of coverage over the area. These referenced plots can be seen in figure 4.23. For the 16th, AOD values are predominantly below 0.5 with a few clustered pixels near the Nevada border reaching above 1. For the 24th, a majority of the AOD values are above 1, especially in the valley regions of California where smoke or aerosol could collect easily.
Using the random.sample provided in Python and choosing a sample size of n=25 from the study region shown above, running the Mann-Whitney U test considering AOD values for the overpasses during the day returned a p-value 4.51010879e-05. Given this value is below 0.05, statistical significance is given to the AOD value.

This AOD alone, however, does not provide information about where the aerosol is in relation to the vertical column, an understanding necessary for comparison to storm cell formation location.

4.5.2 Feature Mask and Vertical Distribution

CALIPSO LiDAR vertical feature mask trends are assessed here to visualize the atmospheric layer most laden with smoke cover. For these LiDAR overpasses, the 15th overpass in figure 4.24 displays the presence of cloud cover and total attenuation as the storms which caused the LIW come into view. The next overpass available, August 18 in figure 4.24 as well, shows an area of cloud cover in the 7.5-10 km range.

Then by August 20 in figure 4.25, a layer of VFM identified aerosol as well as stratospheric aerosol begins to form in the 2-5 km layer. This only becomes more organized in the next August 20 overpass as figure 4.25 also shows a portion of the layer extending above 5 km.

The next overpass in figure 4.26 on August 23 displays this aerosol still in a seemingly more concentrated layer. Then, for the August 25 overpass in figure
Figure 4.24: CALIPSO overpass showing vertical feature mask.

(a) August 15
(b) August 18

Figure 4.25: CALIPSO overpass showing vertical feature mask.

(a) August 20
(b) August 20
4.26 as well, the day after the second case day, a very strong aerosol layer had formed, displaying the overall magnitude of aerosol coverage.

### 4.5.3 Air Quality and Surface

Air quality data coincide with smoke and aerosol coverage as air quality is directly related to smoke concentrations over an area with regards to the plants, animals, and humans living in an area. Air quality data show a sharp decrease in the quality of the air as particulate matter increases during this time are very drastic. The San Jose air quality site shows the amount of change that occurred in air quality during this time, which can be viewed in figure 4.27.

Daily PM 2.5 data are shown throughout the year with the two days of interest noted by colored vertical lines. Prior to this time period, values were consistently below 100 micrograms per cubic meter. However, after this event, maximum values were in excess of 400 micrograms per cubic meter. This is not just for this site, however, as other sites experienced this as well.
Figure 4.27: US EPA air quality data for the San Jose site showing the yearly trend of PM 2.5 over the 2020 period. Days of interest are denoted with vertical lines.
Chapter 5. Discussion and Implications

The motivation for this work is to determine physical and environmental differences between conditions on August 16th and August 24th, 2020, in order to improve future wildfire forecasts.

When viewing the model data, one can see ridging over the western United States with high pressure near the four corners region. The same general pattern is present on both days with more organized high pressure present on the 16th. A low-pressure system to the west and associated troughing moves shoreward as higher values of vorticity enter the area.

Data gathered in this study supports the hypothesis that aerosols potentially altered the cloud formation and lightning formation processes. LCL heights could increase due to scavenging of moisture due to smaller, more numerous CCN present in the area. These smaller and more numerous CCN would steal all the moisture present at the lower levels and not leave any for the cloud formation process.

When considering the correlation matrices, the 24th seems, from a radar’s perspective, as if it were the drier and more prolific fire starter of the two cases studied here. Although there were very little CG making it to the ground, a drier setup would actually increase lightning chances and storm precipitation
under ordinary circumstances. When considering the cells which were present, the reflectivity distributions for these cells were drastically different. The 16th had more cell snapshots which had higher values of reflectivity returns than the 24th had. These also had a 24.5% increase in the number of cells which had VII present. This is notable due to the necessity for ice and graupel coexistence for the formation of lightning. Without these processes, lightning will not be present. The echo tops on the 16th and 24th were different as well. The average 18 dBZ echo top for the 16th was 1.7 km higher than on the 24th while the 30 dBZ echo top and freezing level were 2.2 and 0.09 km different, respectively. There is a stark contrast between the echo tops while the freezing levels do not have as much of a drastic change.

A few theories exist as to why this might be the case. If elevated aerosols are altering cloud properties, these changes could occur in the microphysical aspects of the cloud. The overall ambient temperature of the environment, however, would still be the same if microphysical changes were occurring. So, in addition to overall land conditions, synoptic conditions, and general large-scale considerations being similar, one can see that the ambient atmosphere is also very similar based on the 0.09 km difference in model estimated freezing level height difference. Therefore, one can assume that between the freezing level and 30 dBZ or 18 dBZ echo top, something is occurring within the cloud to alter the overall convective properties as there are stark differences in heights at these levels. Thus, vertical feature mask information is considered.
On the vertical feature mask diagrams, one can see the presence of a very distinct aerosol layer between 3-6 km by the time of the second case day. This layer would not be static, however, as cloud formation and the atmosphere as a whole is a very chaotic system. Constant upward and downward motions, whether it be due to thermals, boundary layer influences, inversions, general atmospheric continuity processes, or other processes, are always occurring. These movements would create movement of the aerosol and, due to entrainment or typical flow processes, the aerosol then becomes mixed as the other air in the atmosphere would be.

There were small differences in freezing level heights. Due to this, some change has to occur between the freezing level and the top of the cloud as the 18 and 30 dBZ echo tops are different. In this area of the cloud, the formation of SCLW and graupel would begin to help create the formation of a mixed-phase zone. As the temperatures reach approximately -10°C, the area of the cloud most prone to the presence of these three phases due to supersaturation processes, lightning potential increases as the mixed-phase matures. At this level, higher reflectivity values signal the presence of more, larger hydrometeors as reflectivity is a size-dominated value. Looking at the ice mass vs. maximum -10°C reflectivity chart helps to better understand this level. Reflectivities above 30 dBZ are often assumed to signal the presence of graupel based on graupel’s typical number concentration and size. Here, the 30 dBZ level is near where the biggest difference in the two days starts to occur. On the 16th, as the reflectivity increases, so does the ice mass. It trends in a near logarithmic pattern. This is not the case on the
24th however, as the ice mass stays well below even 200 million kg while the 16th got up to above 900 million kg.

So, a few ideas emerge out of that. There could be:

more smaller hydrometeors = smaller reflectivities = smaller vii return
less smaller hydrometeors = much smaller reflectivities = much smaller vii
less larger hydrometeors = larger reflectivities = larger vii returns
more larger hydrometeors = much larger reflectivities = much larger vii

Some of the cells possessed more VII than others, and those cells with the larger volumes are expected to also be clouds with larger amounts of lightning as well. To get ice and form a mixed-phase region, those ideal temperatures need to be present. Another value assessed to help in that consideration was the presence of the -10°C level. Several of the 24th cells did not even make it to this level, helping to support why the lightning was not as numerous as the ice mass was very different.

For the cloud to reach those temperatures, there have to be upward motions to encourage lifting. This lifting is in part due to the condensational heating of water on the available CCN present in the current atmosphere. This has to occur above the LCL, however. So, knowing where the LCL is helps in assessing where the cloud base can form. Here, the 24th cloud base was higher for the earlier part of the period, meaning that CCN have to make it to higher heights before condensational heating or scavenging of moisture can occur.

There must be ample moisture and instability for these processes to take place. Each of these systems were associated with post tropical systems, so there
was no shortage of moisture there. These CAPE values, although low compared
to typical thunderstorms of the southeastern U.S., are near or above average for
most of the period. CCN for the moisture to condense on would also be necessary.
Could the aerosol serve as these CCN?

Lightning for August 16th follows common trends as there is typically
a more linear distribution between the ice mass and lightning characteristics of
a cloud, given the ice-lightning relationship (Deierling et al. 2008, Mosier et
al. 2011). The CG vs. VII pattern on the 24th, however, does not represent
this typical pattern as there were many cells which had values of VII but did
not produce CG returns. These values of VII are rather small, but there is a
congregation of these values on the 16th as well. So, given this data, these values
of VII in a typical environment can indeed produce CG lightning. The present
ice mass did not, however, produce the expected amounts of CG lightning on the
24th based on common trends.

This is interesting. VII is a radar-derived value. It estimates the precipita-
tion ice mass within the thermodynamic layer bounded by the -10°C layer and
-40°C layer considering the variables of radar reflectivity and heights of -10°C and
-40°C. Since radar reflectivity considers the number concentration and diameter
of hydrometeors with diameter raised to the 6th, the diameter of a hydrometeor
has a much heavier weight on the reflectivity than the number concentration. So,
the size of hydrometeors would change VII more drastically than the number con-
centration of hydrometeors altogether. Assuming aerosol impacts on a cloud, the
cloud would have smaller hydrometeors with a larger number concentration given
an estimated increase in CCN due to the aerosol increase. There are caveats to this, however, as the aerosol and CCN are not perfectly equal. Assumptions can be made about CCN amount given AOD. It must be noted, however, that CCN and AOD are not one-to-one.

Another aerosol theory essential to consider is the presence of an inversion or direct heating, as mentioned in past literature. This is important due to the fact that there was an inversion present on the August 24 sounding. This inversion, in addition to the little surface moisture present and dry boundary layer aloft, suggest the potential for direct effects of aerosol influencing the August 24 atmosphere as well.

Aerosols also alter convective eddies and upward motions. Based on previous research into the NIC method, there is a direct relationship between upward strength and lightning potential. This can occur due to the thermodynamic differences which were discussed earlier as well. Again, it must be noted that the thermodynamic differences are important and something that cannot be overlooked. Increased buoyancy and CAPE increases upward motions. These upward motions lead to cooler cloud temperatures and increased potential for a mixed-phase region, which is necessary for lightning. Very high concentrations of aerosols could also hinder the lightning potential in this way as the upward motions are suppressed and convective initiation is hindered.

As discussed in the earlier results section, the aerosol difference for each of these cases was a very novel finding of this study. Based on the background knowledge surrounding aerosol and cloud processes, the direct, indirect, and semi-
direct effect of aerosols on clouds could have influenced these results in several different ways.

Many studies in the past have attempted to assess this complicated relationship between aerosol and cloud invigoration/suppression processes. The aerosol and associated processes relationship is a very unsettled one, as many varying factors play a role in these activities. Other studies have also attempted to address and better explain this noticeable difference in findings pertaining to the aerosol-cloud relationship (Khain 2009, Dayeh et al. 2021). Here in table 5.1, several of these processes and the associated pros and cons of lightning production have been summarized to better display how these varying aerosol effects and implications would impact lightning production in general.

Assuming the direct effect of aerosol on the cloud, one would expect to see evidence of direct radiative forcings (Sekiguchi et al. 2003, Papadimas et al. 2012, Podgorny and Ramanathan 2001, Benedetti and Vitart 2018). These forcings could be present in ways such as an upper-level warm layer, cooling at the surface, or decreased surface-level radiation due to scattering and absorption (Koren et al. 2004, 2005). The August 24 inversion could be attributed to the direct effect of aerosol forcings as this inversion is present in the same vertical layer where the aerosol was identified.
Table 5.1: A table identifying general common aerosol processes, associated relevant sources, and the pros and cons when considering an environment and its ability to create lightning in convective storms. Note that the assumptions of the pros/cons in this table are that the environmental conditions (e.g., low-level moisture, convective available potential energy, vertical wind shear) are not impacted by aerosol concentrations.

<table>
<thead>
<tr>
<th>Process/Sources:</th>
<th>Pro:</th>
<th>Con:</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aerosol concentraions / Wang et al. (2018)</td>
<td>Conditions favorable for lightning increase with low aerosol concentrations</td>
<td>Lightning is inhibited with relatively high aerosol concentrations</td>
</tr>
<tr>
<td>Increasing CCN concentrations reduces the radius of cloud droplets, enhanced cloud drop evaporation, increased in-cloud evaporation, decreased buoyancy, negative aerosol effect leads to an enervation of deep convection, decreased precipitation efficiency, and weakened updraft strength / Barthlott et al. (2022), Yuan et al. (2011)</td>
<td>Reduced size of cloud droplets and overall precipitation efficiency results in reduced potential of deep convection, and subsequently deep mixed-phase region, therefore decreased lightning potential</td>
<td></td>
</tr>
</tbody>
</table>
Table 5.1 – continued from previous page

<table>
<thead>
<tr>
<th>Process/Sources:</th>
<th>Pro:</th>
<th>Con:</th>
</tr>
</thead>
<tbody>
<tr>
<td>Aerosols act as cloud condensation nuclei (CCN) - delayed or suppressed warm rain process development, enhanced mixed-phase processes - enables light water droplets to ascend further into the upper layers of the clouds, fueling lightning processes, invigorated deep cloud development / Thornton et al. (2017)</td>
<td>Enhancement of lightning flash rates because of further ascent of water droplet and cloud, and more vigorous updrafts in mixed-phase cloud regions</td>
<td></td>
</tr>
<tr>
<td>Reduced incoming solar radiation, suppressed surface heating, altered production of convective eddies and upward motion / Koren et al. (2004, 2005), Wang et al. (2014)</td>
<td>The lightning would be a lot less likely to occur because of the suppressed surface heating and hindered upward motions</td>
<td></td>
</tr>
</tbody>
</table>

Continued on next page
Process/Sources: Aerosol warms middle and/or upper troposphere which alters vertical atmospheric heating profile impacting stability and the formation of deep convective circulations / Allen et al. (2019)

Pro: The increased heating stabilizes the area and prevents upward motions, hindering lightning potential

Con: The increased heating stabilizes the area and prevents upward motions, hindering lightning potential

Assuming the indirect effect of aerosol on the cloud, one would expect to see differences in the microphysical structure of the cloud (Rosenfeld et al. 2014, Kaufman et al. 2005, Yuan et al. 2011). Although the actual microphysical structure of these clouds is not retrieved in this study, the outcomes of microphysical changes which have been documented in these past studies and others (Twomey 1974, Rosenfeld 2000, Albrecht 1989, Pincus and Baker 1994) can be assessed here given these previous results. Shorter cloud top heights with lesser reflectivities and more widespread cells on the August 24 day support the potential for less lightning as these cells also have less cloud volume above the freezing level. The reduced volume of cloud ice supports the microphysical changes as well as there were differences in the number of hydrometeors within that mixed-phase region.

Assuming the semi-direct effect of aerosol on the cloud, one would expect to see thermodynamic changes assuming all else were equal (Allen et al. 2019,
This is a very complex situation as the natural atmosphere is constantly changing. There were thermodynamic differences for these events; however, based on the data that is currently available, these changes cannot be directly linked to the aerosol semi-direct influence on the cloud.

If the aerosol concentration is less, i.e., lower AOD values, then the cloud is actually invigorated according to some studies (Liu et al. 2021, Thornton et al. 2017, Zhao et al. 2015). However, as the aerosol layer thickens, the increasing thickness of aerosol eventually leads to suppressed convection. The size and concentration of the aerosol coverage is the driving factor in the overall thunderstorm effects of invigoration or suppression and how the aerosol eventually alters the cloud processes. In this case, a suppression mechanism is hypothesized to be occurring where storms with lesser numbers of lightning developed due to the smothering of the storms by the thick aerosol layer and previously discussed effects.

It must also be mentioned that these three aerosol effects can alter the thermodynamic profile themselves based on the past studies of aerosol influences. Therefore, it is extremely hard to separate thermodynamic influences from aerosol influences as the aerosols have also been known to alter the thermodynamic environment.

Based on the background works presented in this study and the observations previously discussed, conceptual models have been formulated estimating the cloud structure given aerosol impacts and the resulting changes in cloud processes which would occur based on the data presented in this study, figures 5.1
and 5.2. When considering the lower cloud heights and decreased depth of the mixed-phase region, the lightning supportive portion of the cloud has a significantly reduced amount of volume considering the impact increased aerosol coverage would have on the cloud. There will be caveats with this as well as current information does not provide a clear distinction between aerosol acting as CCN and aerosol acting as ice nuclei (IN) in the cloud. These simple models assume the impacts of aerosols alone and do not include thermodynamic influences.

Figure 5.1: A conceptual model of the August 16th setup and common processes assumed to impact the day.

For the positive LIW case, the model in figure 5.1 assumes the cloud is dominated by latent heat of freezing and deposition processes, supporting increased lightning potential as the latent heat of condensational portion of the cloud would be decreased due to the warm and dry layer between the cloud and surface. Evaporation and cooling of the air would support downward motion below the cloud base. On the 24th, the model in figure 5.2 assumes much smaller contributions of latent heating of freezing and deposition when compared to the
model 16th cloud as there is a smaller portion of cloud that would be in the mixed-phase region due to reduced cloud top heights and reduced mixed-phase depths. These decreases then create decreased lightning potential as the cloud has similar magnitudes of latent heating of freezing/deposition and latent heating of condensation when compared to the 16th conceptual cloud. This decreases lightning potential, moisture transfer, and energy and energy transfer from the surface back up to the atmosphere.

Figure 5.2: A conceptual model of the August 24th setup and common processes assumed to impact the day.
Chapter 6. Conclusions and Future Work

Aerosols have the ability to and indeed do alter storm-scale processes. August 16 and 24, 2020 were unique events where similar land and synoptic scale conditions were present supporting the chances for extreme lightning-initiated wildfire outbreaks. Despite these similar forecasts, vastly different outcomes occurred.

There were notable differences in the vertical structures of thunderstorms on August 16th and August 24th that were present. Shallower mixed-phase depths and smaller ice masses present on August 24th suggest that it is physically plausible that aerosol load from prior fires can contributed to weaker mixed-phase regions in thunderstorms. Present thermodynamic differences, particularly in convective available potential energy and deep moisture, were present and must be considered in addition to aerosol information, however.

Aerosol influence would create more numerous numbers of hydrometeors in the cloud, scavenging the moisture and making the cloud more susceptible to smaller, more numerous cloud droplet growth. These smaller and more numerous droplets would then try to grow by coalescence and end up in the mixed-phase region. These regions were different depths as discovered in the MRMS single-cell
data. Moving forward, current aerosol information may serve as a valuable fire weather forecasting tool.
Potential areas of future work include:

1.) Use reanalysis data to better understand the thermodynamic environment and its impact on the case outcomes. There will be caveats with this approach as well, however, due to the available amount of observations.

2.) Use an MSMS tracker, or a tracker like the TOBAC tracker, to track directly on MRMS instead of using the NEXRAD radar data and then converting over to MRMS. This would presumably be a more efficient method of cell tracking.

3.) Create a database of LIW positive and null cases. Test this hypothesis on a larger dataset and see if it holds true for more events than just those documented in this study.

4.) Potentially look into other MRMS variables, in addition to the ones used here. See if any other variables have a correlation with the cases or if there may be other variables that can be added to make the forecasting of these events better.

5.) Look at other ways to see aerosol vertical distribution, outside of vertical feature mask, such as backscatter or other methods. These other methods may show some characteristics of the aerosols that were not considered here in this study. This could also include the properties of aerosols and how they relate to CCN/IN.

6.) Use modeling techniques to truly test these hypotheses and see, on a larger scale, if this is more common. Through correspondence with NWS forecasters, this is a valid concern and forecasters along the western US actually use this undocumented theory when issuing fire forecasts already. So, actually docu-
menting these types of events and expanding this study could benefit fire weather forecasters tremendously.

7.) Run a field campaign for support/validation. There are many different ways this could happen. Several are really interesting!
References


detection network tm and applications of cloud-to-ground lightning data by
electric power utilities, 1998.

[22] Cummins, K., and Murphy, M. An overview of lightning locating sys-
tems: History, techniques, and data uses, with an in-depth look at the
u.s. nldn. IEEE Transactions on Electromagnetic Compatibility 51 (2009),
499–518.

[23] Dayeh, M., Farahat, A., Ismail-Aldayeh, H., and Abuelgasim,
A. Effects of aerosols on lightning activity over the arabian peninsula.
Atmospheric Research 261 (10 2021).

[24] Deierling, W., Petersen, W., Latham, J., Ellis, S., and Chris-
tian, H. The relationship between lightning activity and ice fluxes in

[25] Doswell, C. The distinction between large-scale and mesoscale contribu-
tion to severe convection: A case study example. Weather and Forecasting

[26] Dowdy, A., and Mills, G. Atmospheric and fuel moisture characteristics
associated with lightning-attributed fires. Journal of Applied Meteorology
and Climatology 51 (11 2012), 2025–2037.

[27] Dye, J. Early electrification and precipitation development in a small, iso-
1231–1247.

within two regions of charge during initial thunderstorm electrification.
Quarterly Journal of the Royal Meteorological Society 114 (7 1988), 1271–
1290.

[29] Dye, J. E., Jones, J. J., Winn, W. P., Cerni, T. A., Gardiner,
Early electrification and precipitation development in a small, isolated mon-


[52] Li, Z., Rosenfeld, D., and Fan, J. Aerosols and their impact on radiation, clouds, precipitation, and severe weather events, 9 2017.


[93] Soloman, S., Qin, D., Manning, M., Chen, Z., Marquis, M., Averyt, K., Tignor, M., and Miller, H. Contribution of working group i to the fourth assessment report of the intergovernmental panel on climate change.


