

UNIVERSITY OF OKLAHOMA

A Case Study on Diurnal Boundary Layer Evolution

Meteorological Measurement Systems – Fall 2010

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Abstract

In this project, data from four different soundings over Norman, Oklahoma from September 17, 2010 was used to evaluate changes in the boundary layer. The boundary layer was identified by determining the height of an inversion layer and other thermal properties of the low levels of the atmosphere. It was determined that as the day progressed, the boundary layer deepened as the inversion layer rose and low-level lapse rates steepened. The depth of the boundary layer has important applications in forecasting daily weather conditions and severe weather phenomena.

Introduction

For meteorologists, knowing the three-dimensional state of the atmosphere is crucial for both forecasting and research. Surface observations only complete part of the picture; in order to obtain weather observations *above* the surface, upper-air measurements must be taken. Upper-air measurements may be taken in several different ways, but the best way to get a complete picture of the upper-levels of the troposphere (where virtually all weather occurs), is to launch radiosondes and rawinsondes. A radiosonde is a package of meteorological instruments that are launched into the troposphere by a weather balloon (American Meteorological Society Glossary). A special type of radiosonde that can measure wind speed and direction is called a rawinsonde. In this particular project, rawinsondes were used to measure temperature, humidity, pressure, and wind speed and direction.

Errors in the data can exist due to basic instrumental error (response time and resolution) and outside influences. Temperature measurements can be subjected to error from solar radiation, but these errors are minimized by the aluminum coating on the temperature sensor. This coating reduces solar and infrared heating. Errors in humidity reporting come from the hysteresis of the capacitive polymer that is used for measuring humidity (Miloshevich, et. al., 2001). Also, there can be some errors in height reporting due the use of the hypsometric equation (Parsons, et. al., 1984). Because of this, any errors in temperature, moisture, and pressure reporting will result in height reporting errors.

In this project, upper-air data was used to study the evolution of the planetary boundary layer throughout the day. The planetary boundary layer (PBL) is the bottom layer of the

atmosphere that is in contact with the surface. This layer is subject to turbulence and mixing during the day. The boundary layer can be identified by the region of the atmosphere that lies beneath a low-level inversion layer. Knowing the properties of the boundary layer has important applications in forecasting daily weather conditions and severe weather.

Experimental Details and Methods

In this project, two rawinsondes were launched in Norman, Oklahoma. In addition to our own rawinsonde launches, rawinsonde data from the National Weather Service (NWS) was also used. Our launches were performed on September 17, 2010 at 1400 UTC and 1700 UTC. The NWS launches were performed on the same day at 1200 UTC and on September 18, 2010 at 0000 UTC (all launches were on September 17 local time).

For the 1400 and 1700 UTC launches, InterMet iMet-2 rawinsondes were used. Before the launch, a helium-balloon was inflated and the instrument package was attached to the balloon. Helium was used instead of hydrogen since it is not a flammable gas, and is therefore much safer to use. The only safety concern was the high pressure of the helium tank, but this pressure was controlled using a gas regulator. After the balloon and instrument package were released, the data was transmitted back down to the surface by a radio transmitter. A radio receiver was connected to a computer which recorded the data. Table 1 contains a list of the sensors in the iMet-2 rawinsonde package.

Table 1: iMet-2 Sensors

Variable	Sensor	Accuracy	Resolution	Range
Temperature	Bead Thermistor	0.3°C	0.01°C	-90 to 60 °C
Humidity	Capacitive Polymer	5%	1%	0 to 100%
Wind Direction	GPS	5°	1°	---
Wind Speed	GPS	0.2 m/s	0.1 m/s	---

After receiving the data packages, skew-t/log-p diagrams were generated from a program (*Skew-T/Log-P Plotter*) developed by team member Sam Irons. This program plotted the data on a skew-t/log-p diagram as well as calculated the lifting condensation level (LCL), level of free convection (LFC), equilibrium level (EL), convective available potential energy (CAPE),

convective inhibition (CIN), and precipitable water (PW). Descriptions of these parameters are available in Appendix A.

The variables that were initially collected include temperature, virtual temperature (the temperature moist air would have if dry and at the same pressure), relative humidity, pressure, geopotential height, and wind speed and direction. From these variables, vapor pressure, saturation vapor pressure, mixing ratio, and dew point temperature were calculated. The equations for calculating these variables are in Appendix B.

Results and Discussion

Early morning sounding: 1200 UTC 9/17/2010 (National Weather Service)

To study the early morning boundary layer, the National Weather Service sounding from 1200 UTC (7:00 a.m. local time) was analyzed. The skew-t plot for the 1200 UTC data can be seen in Figure 1.

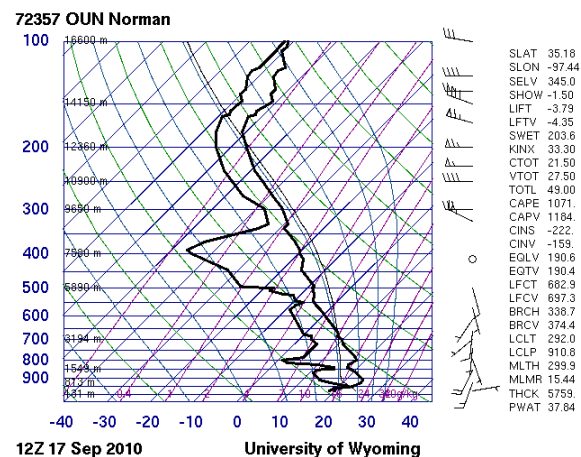


Figure 1: 1200 UTC KOUN Sounding

In Figure 1, it can be observed that from the surface to 900 hPa, an inversion layer is in place. This inversion is the result of boundary layer decoupling that occurred during the overnight hours leading up to this sounding. This boundary layer decoupling is a result of nocturnal radiational cooling. During the overnight hours, there is no incoming solar radiation, but there is outgoing longwave radiation. As longwave radiation is emitted from the Earth, the Earth's surface cools, causing the air closest to the surface to cool the most. This process is most likely to occur on clear, calm nights when there are no clouds to scatter the longwave radiation

back to the Earth's surface and no winds to mix the low-level air (Haby, J., 2010). The low-level nocturnal inversion is very evident in this sounding that was taken just before sunrise (sunrise and sunset data are available in Appendix C).

Late morning sounding: 1400 UTC (iMet-2 launch #1)

The first launch by our team occurred at 1400 UTC (9:00 a.m. local time), two hours after the morning National Weather Service sounding. The skew-t plot for this sounding is shown in Figure 2.

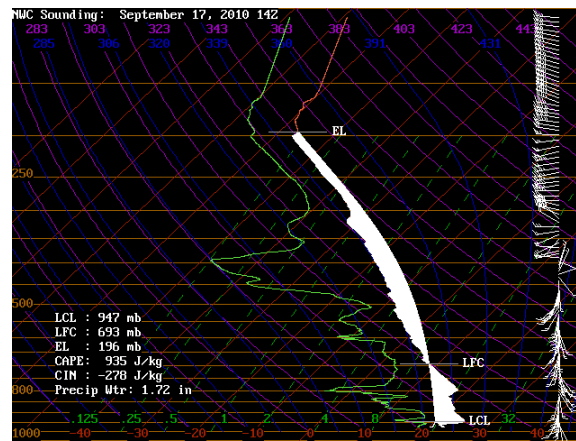


Figure 2: 1400 UTC iMet-2 Sounding

In Figure 2, temperatures are now cooling with height in the lower 10 hPa of the atmosphere. It is evident that there have been some small effects from turbulence and surface heating, just two hours after sunrise. Also note that the LCL is at 947 hPa, the LFC is at 693 hPa, CAPE is 935 J kg^{-1} , and CIN is -278 J kg^{-1} ; these values will change throughout the day as heating processes continue.

Afternoon Sounding: 1700 UTC (iMet-2 launch #2)

Our second launch occurred at 1700 UTC (12:00 p.m. local time). The skew-t plot for this launch can be found in Figure 3.

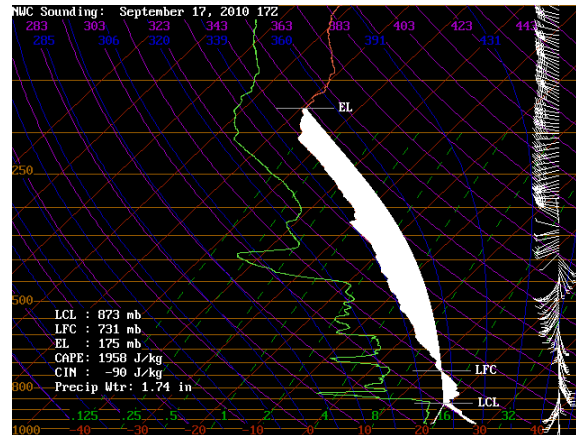


Figure 3: 1700 UTC iMet-2 Sounding

In Figure 3, it can be observed that the temperatures in the lowest 120 hPa of the atmosphere are now cooling with height. With five hours of surface heating, thermal and mechanical turbulence have begun to mix the boundary layer very well. Two observations can be made to determine that the boundary layer is beginning to mix. First, the low-level lapse rate (change in temperature with height) is now down to $-7.5^{\circ}\text{C km}^{-1}$ and is approaching the dry adiabatic lapse rate of $-9.8^{\circ}\text{C km}^{-1}$. Secondly, the dew point line is nearly parallel to the isohumes (dashed green lines on the skew-t). This shows the mixing ratio is being conserved in the boundary layer, which is evidence that moisture is being well-mixed in the lowest levels of the atmosphere.

From the skew-t parameters, it can be determined that the LCL has increased in height from 947 hPa to 873 hPa in just three hours. Due to a one degree rise in dew point temperature, but a five degree rise in temperature, the dew point depression has increased. Because of this increase in dew point depression, parcels must be lifted higher in order to achieve saturation, thus the increase in LCL height. While the LCL rose, the LFC fell from 693 hPa to 731 hPa. This lowering of the LFC is due to steepened lapse rates at the lower levels of the atmosphere. There is also an increase in instability and decrease in cap strength due to the lowering of the LFC; it takes *less* energy to raise parcels to the LFC and there is more depth between the LFC and the

EL. There was little change in the height of the EL since it lies well above the boundary layer is not influenced by thermals and mechanical turbulence.

Another noticeable feature is the similarity of the winds between the morning and afternoon soundings. This lack of a significant change is due to weak wind speeds aloft. The main reason winds tend to increase during the day is the downward transfer of linear momentum in the atmosphere. Higher winds aloft would “mix down” to the lower levels, resulting in an increase in low-level winds, but on this particular day, winds were weak aloft with wind speeds of only 14 knots at 500 hPa. This can be contrasted to a typical severe weather day where 500 hPa winds can exceed 50 knots. A plot of winds from the 1400 UTC and 1700 UTC soundings can be found in Figure 4. In this figure, the blue line indicates morning wind speeds (1400 UTC) and the red line indicates afternoon wind speeds (1700 UTC).

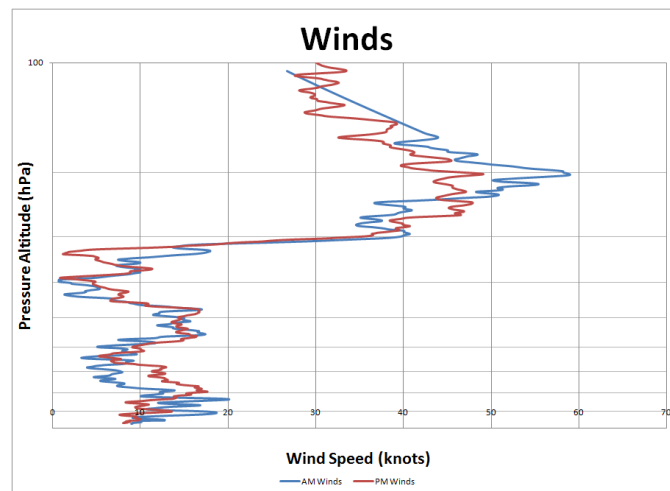


Figure 4: Wind Speeds from the iMet-2 launches

Evening sounding: 0000 UTC 9/18/2010 (NWS Sounding)

The last sounding used in this case study was the 0000 UTC 9/18 (7:00 p.m. 9/17 local time) sounding from the National Weather Service. The skew-t plot for this sounding is in Figure 5.

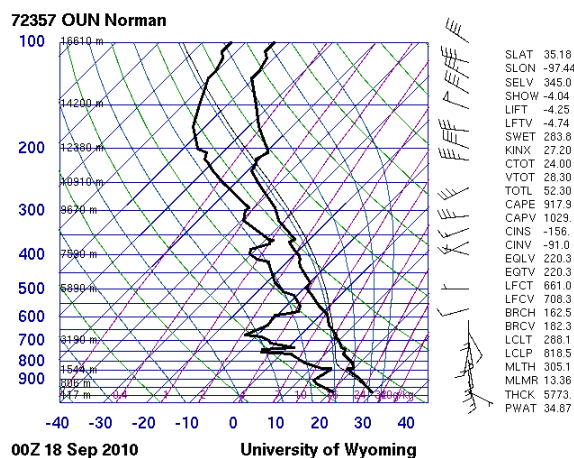


Figure 5: 0000 UTC KOUN Sounding

In the final sounding, the boundary layer extended throughout the bottom 140 hPa of the atmosphere. The inversion layer is now located at about 840 hPa with a low-level lapse rate of about $-8.3^{\circ}\text{C km}^{-1}$. The most interesting feature on this sounding is the “dew point inversion” located at around 900 hPa. This lowering of the dew point with height in the lowest levels of the atmosphere is due to moisture being mixed upwards. Above the inversion, dew point temperatures increase up to 850 hPa. For the portion the dew point temperature increases, the dew point curve is still nearly parallel to the mixing ratio line, indicating constant moisture content with height up to this level.

The trend of increasing LCL height with time also continued from 1700 UTC to 0000 UTC due to the small change in surface dew point, but continued increase of the surface temperature. Also during this time, the LFC height increased. This increase in height is likely correlated with continued reduction in surface relative humidity. Since parcels must be lifted even higher than before to become saturated, they do not begin rising moist adiabatically until reaching a higher level. This means the parcel temperature does not become warmer than the environment (which has also warmed throughout the day) until the parcels reach a higher level. This increase in LFC height, but continued little change in EL height, has resulted in a net

decrease in CAPE from 1700 UTC to 0000 UTC. While CAPE decreased, CIN has increased once again due to the increase in LFC height.

Conclusions and Summary

As the day progressed, the boundary layer increased in depth due to surface heating (thermal eddies) and weak turbulence, with the largest increase in depth occurring between 1400 UTC and 1700 UTC. This boundary layer depth increase was marked by an increase in height of the inversion layer and steepening low-level lapse rates throughout the day. Due to the nearly constant moisture content of the low levels, and increase in surface temperatures, the relative humidity at the surface decreased, thus raising the LCL height. The precipitable water (a measure of the total moisture content of the atmospheric profile) stayed nearly constant throughout the day, indicating that the moisture present later in the day was the same moisture that was present at the beginning of the day and that the moisture was transferred vertically through mixing, and not horizontally through advection. The steepening lapse rates initially resulted in a drop in the LFC height during the early afternoon hours, but the LFC level rose later in the afternoon as the temperature of the atmospheric profile increased throughout the entire atmosphere. The likely reason that the low-level temperatures rose faster than the upper-level temperatures is that the ground temperature increases faster due to the lower specific heat of the ground compared to the air. Since the ground heats faster than the air, heat is transferred from the ground into the lower-levels of the atmosphere, thus creating the “lag time” in temperature change between the low levels and upper levels.

In conclusion, the boundary layer was able to be identified based on thermal profiles of the lowest levels of the atmosphere, particularly the height of the inversion layer. As the day progressed, low level lapse rates increased and the height of the inversion layer was forced higher. On this particular day, there was no deep convection due to the persistent inversion layer, at first near the surface then later around 840 hPa. On a typical severe weather day, there is likely to be a weaker surface inversion in the morning that weakens during the day and completely disappears by mid afternoon when thunderstorms initiate. For example, in Figure 6, a set of soundings from the May 19, 2010 severe weather event in central Oklahoma, there is initially a strong inversion in the morning (leftmost sounding) that weakens throughout the day to the point where it is virtually non-existent in the low levels of the atmosphere by late afternoon (rightmost

sounding). It can also be seen that the low-level lapse rates steepen throughout the day. The only way deep convection could be initiated on a day similar to September 17, 2010, would be if a frontal system (or other source of lift) approached the region. Later in the afternoon, the atmosphere was fairly unstable above the boundary layer, meaning if a cold front, dry line, or other lifting mechanism were in the area, deep convection could have been initiated if parcels were able to be forced above the boundary layer.

Frequent, high resolution soundings can help meteorologists better understand boundary layer conditions. Knowing the state of the boundary layer is very important in forecasting. Boundary layer decoupling during the overnight hours can be critical in forecasting overnight low temperatures and fog development. Boundary layer meteorology also plays an important role in severe weather forecasting. The thermal conditions of the lowest level of the atmosphere are crucial for forecasting convective initiation and the severity of thunderstorms, particularly the existence of thermal inversions and steep low-level lapse rates. Boundary layer meteorology continues to be an important part of meteorological research as we strive to better understand these processes.

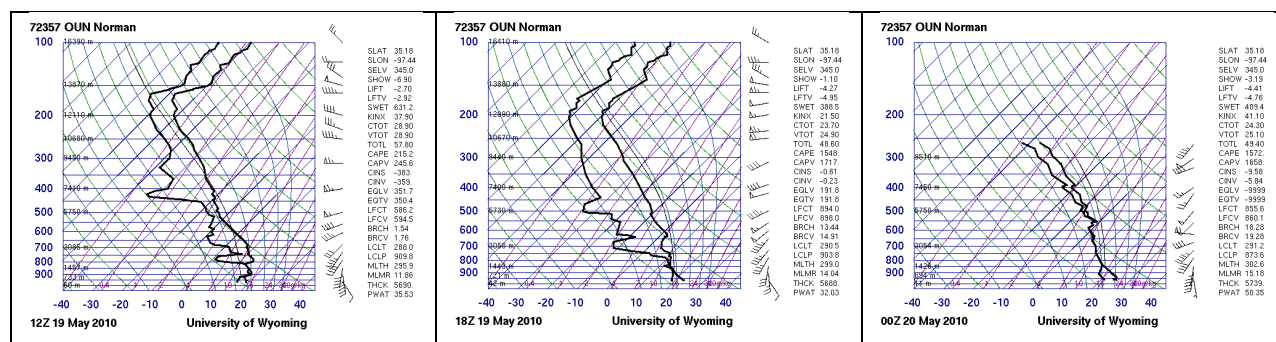


Figure 6: KOUN Soundings from May 19, 2010 Severe Weather Event (from left to right, 19/12Z, 19/18Z, 20/00Z)

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Appendices

Appendix A: Skew-T/Log-P Parameters

Lifting Condensation Level (LCL): the LCL is the height at which a parcel of air becomes saturated when lifted dry adiabatically. The LCL can be found by finding the surface dew point temperature and surface temperature, then finding the intersection of the mixing ratio line that the dew point temperature lies on and the dry adiabat that the temperature lies on. The level of this intersection is the LCL.

Level of Free Convection (LFC): the LFC is the height at which a parcel lifted dry adiabatically to the LCL then moist adiabatically becomes warmer than the environment and thus less dense, allowing it to rise freely until the parcel is cooler than the surrounding air.

Equilibrium Level (EL): the EL is the height at which the temperature of a rising parcel becomes equal to the environmental temperature and thus stops rising freely.

Convective Available Potential Energy (CAPE): the maximum amount of energy available to an ascending parcel, calculated by computing the area on a skew-t between the environmental and parcel temperatures, from the LFC to EL.

Convective Inhibition (CIN): the amount of energy required to lift a parcel to the LFC, calculated by computing the area on a skew-t (in the region where the environmental temperature is warmer than the parcel temperature), below from the parcel origin (typically the surface) to the LFC.

Precipitable Water (PW): the amount of water that would be collected over a cross-sectional area if all water vapor in a column of the same cross-sectional area were condensed and precipitated out.

Appendix B

Saturation vapor pressure (e_s): the vapor pressure at a given temperature at which the vapor of a substance is in equilibrium. The saturation vapor pressure is calculated using the Clausius-Clapeyron Equation, given by Equation 1; where e_0 is 6.11 hPa, T_0 is 273 K, and T is temperature in Kelvins. Multiplying this value by relative humidity will yield the vapor pressure (e). By substituting in vapor pressure for saturation vapor pressure and solving for temperature, we can develop an equation for dew point temperature (T_d), the temperature at which air at constant pressure must be cooled in order to reach saturation. The equation for dew point is given by Equation 2.

$$e_s = e_0 \exp \left[\frac{L}{R_v} \left(\frac{1}{T_0} - \frac{1}{T} \right) \right]$$

Equation 1: Clausius-Clapeyron Equation

$$T_d = - \left[\ln \left(\frac{e}{e_0} \right) \frac{R_v}{L} - \frac{1}{T_0} \right]^{-1}$$

Equation 2: Dew Point Temperature

Mixing ratio (w): the dimensionless ratio of the mass of water vapor to mass of dry air, given by Equation 3; where ϵ is 0.622, e is vapor pressure in hectopascals and p is pressure in hectopascals.

$$w = \frac{\epsilon e}{p - e}$$

Equation 3: Mixing Ratio

Appendix C: Sunrise and Sunset

This table shows the sunrise and sunset times for Norman, Oklahoma on September 17, 2010. Source: Naval Meteorology and Oceanography Command

Sunrise	Sunset
7:14 a.m. (1214 UTC)	7:34 p.m. (0034 UTC)