Sudden Nocturnal Warming Events in Mississippi

LOREN D. WHITE
Jackson State University, Jackson, Mississippi

(Manuscript received 25 February 2008, in final form 16 September 2008)

ABSTRACT

Significant episodes of sudden nocturnal warming have been observed by the Mississippi Mesonet. The probable relation of these nocturnal warming events to surface layer regime transitions between a decoupled quiescent surface layer and a more turbulent, less thermodynamically stable surface layer is discussed within the context of four examples with different temporal signatures. In general, the changes in wind speed and inversion strength are consistent with expectations for such regime changes. However, details of individual events indicate a wider variety of event characteristics than has been documented previously. The cases examined are proposed as prototypes for four different types of warming event, based on the evolution of temperature and dewpoint as well as on whether clear forcing from a mesoscale or synoptic frontal passage can be identified. Using this classification system and a subjective evaluation of event magnitude, the frequency of nocturnal warming events is analyzed for four mesonet stations at varying distance inland over the period of record.

1. Introduction

Regime transitions in the nocturnal atmospheric boundary layer have in general been less well documented than daytime features, in particular for environments of weak atmospheric and topographic forcing. Under clear-sky conditions, development of strong thermal stability due to surface radiative cooling tends to strongly inhibit turbulent motions at night (e.g., Poulos et al. 2002). This may result in a decoupling of the surface layer from the free atmosphere, with a resulting decrease in wind velocity to near-calm conditions. Hence, in many areas of the southeastern United States (away from coastlines or steep topography) the typical near-surface meteorological behavior at night under weak synoptic/mesoscale forcing may be idealized by weak winds and a gradual decrease in temperature to approach the dewpoint. In addition, the evolution of humidity is normally influenced by the evening transition, as described in detail by Mahrt (1981). This phenomenon may be summarized as an accumulation of water vapor in the surface layer during the first few hours after sunset due to the sudden regime change that often occurs from an unstable daytime convective profile to the stable nocturnal profile. It is most clearly manifested in time series by an increase in dewpoint (Fig. 1) or other moisture variable. The evening transition and a similar rise in humidity shortly after sunrise have been discussed in various contexts for several decades (e.g., Geiger 1950).

On 1 April 2004 the first station of the Mississippi Mesonet (White and Matlack 2007) began recording data at Newton, at 1-min intervals. Upon examination of the data, major deviations from the normal pattern of nocturnal cooling were noted on some nights. Cooling would occur at a fairly steady rate for some period after sunset, and then the temperature would suddenly rise significantly. Dramatic changes of other observed parameters (such as dewpoint and wind speed) were correlated with the temperature changes. As other stations were installed, similar behavior was sometimes observed. Because most of these events did not seem to be associated with warm fronts or heat bursts (Johnson et al. 1989) and the mechanisms were uncertain, they were generically categorized as “nocturnal warming events.”

Similar events can be noted on 5 and 29 October 1982 in data gathered near Albany, New York, by Fitzjarrald and Lala (1989) [and later discussed by Acevedo and Fitzjarrald (2001)], although their discussion was...
primarily concerned with the humidity increase during the rapid cooling of the evening transition. Recent field campaigns such as the Cooperative Atmosphere–Surface Exchange Study 1999 (CASES-99) have investigated nocturnal boundary layer phenomena, including the intermittent turbulence often associated with nocturnal warming (LeMone et al. 2000; Poulos et al. 2002).

Several possible mechanisms for the nocturnal warming events observed in Mississippi have been considered, with the possibility that multiple mechanisms could be responsible. The theoretical implications for consideration of most of these hypotheses will be discussed, following a description of specific cases with different characteristics. For now it will suffice to list potential mechanisms for nocturnal warming:

1) warm-frontal passage,
2) heat bursts driven by decaying convective storms,
3) turbulent overturning or mixing of the nocturnal radiation inversion due to increased wind speed,
4) change in radiative balance due to change from clear to cloudy skies, and
5) latent heat release from fog or dew formation.

Adiabatic warming by katabatic winds is irrelevant because of the lack of sufficient terrain in the region. Although nocturnal drainage winds can occur within the region, the elevation difference would be insufficient for significant warming.

It became clear that many nocturnal warming events observed by the Mississippi Mesonet could not be explained by warm fronts or heat bursts because there were no synoptic-scale fronts or previous convective activity in the area. In addition, note that heat bursts (Fiebrich and Crawford 2001) have yet to be documented in the Southeast, with almost all documented cases being from the Great Plains. In the typical case, heat bursts are accompanied by strong (potentially damaging) winds, which are not characteristic of most nocturnal warming events observed in Mississippi.

Mixing by increased turbulence was attributed as the cause of the warming cases of Acevedo and Fitzjarrald (2001). Such “breakdowns” of static stability were documented as common features of the nocturnal boundary layer by Nappo (1991). Reina and Mahrt (2005) and Van de Wiel et al. (2003) have also documented intermittent turbulent episodes in the stable nocturnal boundary layer during CASES-99. Similar episodes of warming had been noted in Europe by Geiger (1950).

The potential significance to understanding nocturnal warming events is threefold. First, the sudden, almost discontinuous nature of some nocturnal warming events may hold keys to improved understanding of regime changes in the nocturnal boundary layer in general, with implications for fumigation of pollutants (Weber and Kurzeja 1991) and parameterization of boundary layer processes for numerical models (Derbyshire 1999). A related concern is the impact of such episodic events on the budget of carbon dioxide for climate and ecosystem studies (Acevedo et al. 2006). Second, there is the possibility that nocturnal warming events may result in failed forecasts of daily minimum temperature, with potential consequences for energy management and agriculture sectors. Third, because relative humidity often decreases dramatically during a warming event,
it is to be expected that there could in some cases be significant temporary improvements in visibility that could be of value to safety of aviation and surface transportation.

Section 2 summarizes the data utilized from the Mississippi Mesonet. Metrics for quantitative description of events are explained in section 3. Four case studies are covered in section 4. In section 5, a broad classification scheme is proposed and some preliminary climatological results are described. Section 6 will address implications for the role of potential mechanisms for initiating sudden nocturnal warming.

2. Data sources

Most data referenced herein have been recorded by sensors of the Mississippi Mesonet (White and Matlack 2007). The network currently consists of eight stations distributed from south to north across Mississippi (Fig. 2). Data from the first four stations installed will be used here. Relevant measurements may be generally summarized as 1) temperature and relative humidity at two vertical levels (from Vaisala, Inc., HMP45C sensors), 2) wind speed and direction at two vertical levels (from R. M. Young Co. wind monitors), and 3) radiative surface temperature (from Apogee, Inc., IRTS-P). All data are recorded at 1-min intervals and are transmitted to a server computer in near–real time. A few periods have missing data due to maintenance issues or data flagged as questionable, slightly shrinking the dataset considered for this study. All times are referenced to central standard time (CST), which is the native time of the network.

The Newton observing site is unobstructed for about 500 m or more except to the east, where there are trees in a low-lying area about 100 m away (Fig. 3a). The site is on a slight southeast–northwest ridge, with local relief of a few meters. During the summer of 2006, there was also a temporary flux station operated by Jackson State University about 300 m to the northwest of the mesonet station. The Calhoun City site is in a flat low-lying area (Fig. 3b) at the southwest edge of a small town (population of fewer than 2000). Topographic variations within 1 km are less than 10 m. A line of trees is located about 40 m to the west, and the nearest buildings are about 100 m to the northeast. The Pascagoula site is in the middle of a flat recreational field (Fig. 3c), approximately 2 km from the Mississippi Sound of the Gulf of Mexico. Residential development surrounds it at a distance of slightly over 100 m in all directions. At Agricola there are trees about 40 m to the south and 100 m to the west (Fig. 3d). The terrain is fairly level, with one-story school buildings located about 70 m to the northeast.

Even though none of the sites has an unobstructed wind fetch of more than 1 km, all are nevertheless reasonably representative of the heterogeneous landscape of the Southeast.

Data and charts from the surface synoptic observing network are referenced for consideration of the large-scale environment. Additional station records archived by the National Oceanic and Atmospheric Administration (NOAA) Meteorological Assimilation Data Ingest System (MADIS; MacDermaid et al. 2005) are also used where appropriate. These include Automatic Position Reporting System as a Weather Network (APRSWXNET; http://www.findu.com/aprswxnet.html) and Remote Automated Weather Station network (Zachariassen et al. 2003) stations. In one case, additional data will also be considered from a Climate Reference Network station (Gallo 2005).

3. Metrics for describing events

Once an event has been subjectively identified, three time points within the event are objectively determined: 1) an “onset” time $t_0$ at which the temperature rise begins, 2) a “peak” time $t_1$ at which the rapid warming is completed, and 3) a “return” time $t_{end}$ at which the temperature has cooled back down to pre-event temperatures. These times define a preonset period, a warming period, and a level-off/return (i.e., postwarming) period.
FIG. 3. The U.S. Geological Survey topographic maps and aerial imagery of Mississippi Mesonet sites referred to in this study (adapted from http://terraserver-usa.com). Elevations are in feet (1 ft $\approx$ 30.5 cm). (a) Newton (contour interval of 10 ft); (b) Calhoun City (primary contour interval of 20 ft; supplementary contour at 5 ft); (c) Pascagoula (contour interval of 5 ft); (d) Agricola (contour interval of 10 ft).
The length of the event is defined as $t_{end} - t_0$. In some cases, the return time may be undefined if sunrise occurs before the temperature cools back to previous levels. To focus only on events with rapid warming, a constraint is introduced that $t_1 - t_0$ (at 2 m) be less than 60 min. Cases considered here have also been screened to not have any measurable precipitation within 3 h of the event.

Temperature change, dewpoint (or other moisture variable) change, mean wind speed, and temperature lapse rate are considered for each time period of an event. Humidity is primarily described in terms of dewpoint and relative humidity to better focus on impacts in a nocturnal surface layer that is often very close to saturation. Although not all sensors had been installed yet, the strong nocturnal warming at Newton on 5 April 2004 was the original event to be noticed in the mesonet data (Fig. 4). Between 0115 and 0142 CST, the 2-m temperature rose 3.4°C. The temperature did not return to prewarming levels until 0421 CST, so that the normal nocturnal cooling was interrupted for a period of about 3 h. Throughout the event, the time series of dewpoint was almost a mirror image of the temperature. Nearby Automated Surface Observation Stations (ASOS) and satellite data indicated clear skies.

4. Examples of nocturnal warming

Characteristics and potential mechanisms will be discussed for four strong cases of nocturnal warming that had significantly different evolution. Although these cases are useful to demonstrate the wide variety of nocturnal warming events, it is not intended that they should necessarily be considered as typical or clearly related to specific unique mechanisms. In comparison with events previously discussed in the literature, the warming in each case was relatively long lasting.

a. Case 1: Sudden warming followed by sudden cooling

On 1 July 2006, a strong nocturnal warming event was observed at Newton (Fig. 5) that exhibited relatively clear relationships among observed variables. The synoptic environment was dominated by a large subtropical ridge producing weak flow off the Gulf of Mexico (Fig. 6). Hourly observations from ASOS stations at Jackson (KJAN; 93 km west of Newton) and Meridian (KMEI; 32 km east) indicated clear skies below 3700 m (12,000 ft) and visibility of 13 km (8 statute mi) or more. Nevertheless, satellite and radar data did show midlevel clouds and scattered decaying convection in the area. Observed winds were light (less than 2 m s$^{-1}$) and variable during the period. No warming was observed by the ASOS sensors. This is consistent with the isolated nature of such events as described in the Acevedo and Fitzjarrald (2001) study. An examination of hourly observations from all observing networks in the area that are archived in MADIS did not show evidence of warming and drying except for one station at Canton (APRSWXNET station CW3691), about 90 km to the west-northwest.

Between 2000 and 2011 CST, the 2-m temperature at Newton rose 2.9°C (Fig. 7a). The warming at 10 m began at 1954 CST and ended at 2010 CST, with a 2.6°C temperature rise. (Sunset was at 1907 CST.) The resulting heating rates are 16.0°C h$^{-1}$ at 2 m and 9.5°C h$^{-1}$ at 10 m. The temperature returned to prewarming levels after a little over 1 h. After the dramatic warming, the temperature was nearly constant until about 2100 CST, when it rapidly began cooling at a rate very similar to the rapid warming. Similar to the behavior in the 5 April 2004 case, the dewpoint closely mirrors the temperature, with a rapid drop of several degrees, nearly constant conditions, and then a rapid rise. This resulted in a drop of relative humidity of about 20% during the warming (Fig. 7b) and a decrease of specific humidity from about 13.4 to 11.4 g kg$^{-1}$. The correlation between temperature and dewpoint at 2 m over the full length of the event was −0.82. There had previously been a slight warming (and drying) episode at 10 m only, just after sunset.

Figure 7c shows the time series of temperature difference between 10 and 2 m, indicating that the strength of the near-surface inversion was reduced by a factor of 2 following the warming. During the hour before the onset at 10 m, the average inversion strength was 2.6°C. The average temperature difference over the hour following the rapid warming was only 1.1°C. The onset of the warming event coincided with the timing of the infrared surface temperature becoming less than the 2-m air temperature, although this may just reflect the slower response of the surface to heating by turbulent mixing.

To quantify wind speed variations (Fig. 7d), mean wind speed was calculated during the 1 h before onset and during the warming. At 2 m, the preonset mean wind speed was only 0.1 m s$^{-1}$, whereas it was 1.8 m s$^{-1}$ during the warming. Corresponding values at 10 m were 0.4 and 2.4 m s$^{-1}$, with a maximum gust of 5.1 m s$^{-1}$. This demonstrates that in a relative sense there was a large increase in wind speed during the warming event, even though the absolute magnitude of the winds remained fairly small throughout. There was only a slight shift in wind direction from south-southeast to south-southwest as the wind speed increased (Fig. 7e).

Two other automated weather stations in the immediate vicinity of the mesonet site recorded data during the event. A Climate Reference Network station (Newton 5 ENE) located about 1.5 km to the northeast
recorded similar warming and modest wind speed increases. At the flux station operated by Jackson State University (300 m northwest) the 30-min mean temperature at 2.3 m increased from 25.9°C before 2000 CST to 28.1°C after 2000 CST. Mean wind speed increased from 0.6 to 1.7 m s\(^{-1}\). The latent heat flux from the 10-Hz eddy covariance system [corrected for density effects using the method of Webb et al. (1980)] was only 1 W m\(^{-2}\) prior to the warming but increased to 31 W m\(^{-2}\) for the period of 2000–2030 CST. Sensible heat fluxes changed from \(26\) to \(39\) W m\(^{-2}\). These nearby measurements are consistent with a horizontal scale for the event of at least a few kilometers and with the hypothesis that warm, dry air was rapidly mixed downward by turbulent eddies.

The fairly rapid transition between boundary layer regimes during the warming and cooling appears almost to represent a hysteresis between two solutions with different degrees of decoupling from the free atmosphere, similar to the cases of Acevedo and Fitzjarrald (2001) and Van de Wiel et al. (2003). The surface layer is tending toward either a nearly constant temperature (over time) with weak winds or toward radiative cooling with nearly calm winds. Because data are not available from above the surface inversion, it is not possible to directly consider acceleration of the flow above due to surface layer decoupling. Such acceleration could decrease the Richardson number sufficiently at the inversion top to initiate turbulent mixing.

b. Case 2: Sudden warming followed by gradual cooling

A case from the morning of 3 March 2006 at Calhoun City (Fig. 8) demonstrates some characteristics that are similar to those of case 1. However, there is a much more gradual cooling period, followed by sunrise at 0623 CST. Synoptic observations (Fig. 9) from surrounding ASOS sites mostly indicated clear skies below 3700 m, except for midlevel clouds around midnight at some sites. Infrared Geostationary Operational Environmental Satellite imagery showed a band of mid- to upper-level clouds that advanced across the region during the event. Surface winds were initially weak but became gusty during the early morning hours. The maximum gust observed was 12 m s\(^{-1}\) at Greenwood (KGWO; 82 km west-southwest) between 0353 and 0453 CST. Initially, calm to light northwest winds predominated across central Mississippi, with stronger northerly winds to the north. Light rain was recorded at Tupelo (KTUP; 69 km northeast) around midnight CST. Warming of 1\(^\circ\)–3°C was observed at most sites, showing that this was not purely a local event. About 16 h before the onset of warming at Calhoun City, a cold-frontal passage had initiated a drop in dewpoint from about 14\(^\circ\) to –4°C (specific humidity from 10 to 3 g kg\(^{-1}\)) and a wind shift from west to north-northwest. However, no boundaries were analyzed in the area by the National Centers for Environmental Prediction Hydrological Prediction Center during the warming event, and observations did not indicate a well-defined spatial pattern of temperature or dewpoint for advective effects. There was no convection in the region to support the possibility of a heat burst. Data from the NOAA Profiler Network wind profiler at Okolona (50 km northeast; not shown) indicated that a wind speed maximum initially detected about 8000 m above mean sea level (MSL) at 2100 CST 2 March had descended to about
2000 m MSL by 0000 CST 3 March. The maximum increase of wind speed was about 10 m s$^{-1}$ at 5000 m. There was relatively little change indicated for winds below 2000 m and little variation of wind direction, although changes in the relatively light boundary layer winds would be relatively difficult for the profiler to detect. In light of the wind profiler and ASOS observations, it seems relatively likely that this case may have been at least indirectly forced by an upper-tropospheric wave disturbance, not associated with a surface frontal boundary.

The primary warming event was preceded by an earlier warming event that was strongest at 10 m (Fig. 10a). The total warming in the main episode was 3.2°C at 2 m and 1.5°C at 10 m, with corresponding heating rates of 12.1°C h$^{-1}$ at 2 m and 4.1°C h$^{-1}$ at 10 m. Because the 10-m level had already warmed substantially during the earlier
event, probably the inversion above 10 m had already been weakened. The time to return to prewarming temperatures was 4.1 h at 2 m but only 2.7 h at 10 m. However, if measured from the onset of the earlier event (at 10 m), then the corresponding times are 5.3 h at 2 m and 4.2 h at 10 m. Because dewpoint did not increase during the temperature recovery, the relative humidity (Fig. 10b) remained low after about a 35% drop (at 2 m) during the warming. Over the full length of the event at 2 m, dewpoint was only weakly anticorrelated with temperature ($-0.51$), whereas during the actual warming phase the correlation was $-0.89$. Inversion strength changed dramatically from 3.5°C to less than 1°C (Fig. 10c) and never reestablished its former strength.

Wind speed gradually increased from the onset until about 0245 CST, after which a speed of around 5 m s$^{-1}$ was maintained at 10 m (Fig. 10d). During the warming period, the 2-m wind speed only averaged 0.9 m s$^{-1}$, although this was still an increase over the previous hour’s mean wind of only 0.3 m s$^{-1}$. At 10 m, the antecedent wind was 0.6 m s$^{-1}$, increasing to 1.3 m s$^{-1}$ during warming. Winds may have been affected somewhat by local obstacles. The onset of warming with such relatively weak wind might indicate a relatively shallow inversion layer, especially in light of the earlier warming at 10 m. Wind direction was consistently northerly at Calhoun City during the primary event, although it had shifted to northwest during the earlier event at 10 m (Fig. 10e). The long period before cooling back to antecedent conditions and the extended period of strong wind may be indicative of attainment of a new equilibrium characterized by a more coupled and turbulent surface layer associated with lower Richardson number.
c. Case 3: Warming with moistening

In some cases the dewpoint does not mirror the temperature but is either nearly constant over time or is positively correlated with the temperature. Many such cases can be attributed to frontal passage or convective systems. However, there still remain some cases without a clear connection to known weather systems. One such case was recorded at Pascagoula on 27–28 February 2006, with steadily increasing dewpoint (Fig. 11). The synoptic environment was one of strong high pressure in a relatively cold, dry air mass (Fig. 12). Modification of the arctic air mass by sensible and latent heat fluxes over the Gulf of Mexico provides at least a partial explanation for the persistently increasing dewpoint during the event. Only a small temperature rise was noted at Mobile (KMOB; 45 km northeast), although a 3°C warming was noted at Gulfport (KGPT; 54 km west). Skies were clear at all ASOS sites, and winds were calm or southerly except for one observation of northerly flow at KMOB at 0300 CST. Mist (under clear skies) was observed at the Pascagoula mesonet site around 0200 CST. Of interest is that the surface layer moistens at a fairly constant rate, regardless of regime, possibly due to the contribution of local evaporation.

There was a large warming at the Pascagoula mesonet site around 2000 CST. However, the dewpoint maintained a steady trend of increasing by about 0.8°C h⁻¹. Only when the temperature began to decrease again was there any decrease of dewpoint. Otherwise, the temperature variations were very similar to case 1, with the largest warming being at 2 m (Fig. 13a), a period of relatively constant temperature, and then a sudden decrease of temperature. At 2 m the temperature increased 4.0°C over a 47-min period, whereas there was only a 1.6°C warming at 10 m. The warming at 10 m began about 15 min earlier than at 2 m but ended at about the same time at both levels. The time to return to antecedent temperature conditions was 6.8 h at 2 m and 6.6 h at 10 m. Despite the rising dewpoint, the warming was sufficient to produce a 15% drop in relative humidity at 2 m (Fig. 13b), with a much smaller decrease at 10 m. Once the inversion was destroyed, there was little variation of relative humidity, either in time or in the vertical direction. The correlation between temperature and dewpoint at 2 m was positive in this case, with a value of 0.30 over the length of the event and 0.57 during the warming period. The average inversion strength for the hour preceding onset was 1.5°C (Fig. 13c). Following the rapid warming period the inversion is completely eliminated, yielding a mean temperature difference of −0.1°C between 10 and 2 m. As the temperatures drop again at the end of the event, the inversion is rapidly reestablished to approximately antecedent conditions.

FIG. 8. Time series of 2-m temperature (solid line) and dewpoint (dotted line) at Calhoun City on 3 Mar 2006. Shading indicates a warming event; \( t_0 \), \( t_1 \), and \( t_{end} \) are as in Fig. 4.
As expected, there was a large relative increase in wind speed (Fig. 13d) during the event, first appearing at 10 m. At 2 m the mean antecedent wind speed was less than 0.1 m s\(^{-1}\), whereas it was 0.3 m s\(^{-1}\) at 10 m. Corresponding wind speeds during the warming were 0.6 and 1.2 m s\(^{-1}\). Throughout the event, winds were fairly gusty, with peak speeds exceeding 4 m s\(^{-1}\). Around 0200 CST the wind at both levels decreased to a relatively weak 0.5 m s\(^{-1}\). This corresponds not only to reestablishment of the inversion by radiative cooling but to a wind shift to northwest (Fig. 13e). Winds from the south-southwest had prevailed up to this point. Because the geostrophic wind was weakly from the south and there was no nearby convective or frontal activity, it appears that this wind shift was the delayed onset of the land breeze. It is presumed that the land–sea temperature contrast was not sufficient to force the land breeze until after radiational cooling had been reestablished long enough to drop temperatures back to prewarming conditions. Although the warming event clearly occurred in an advective environment, all of the sudden warming occurred significantly before the change of wind direction associated with the land breeze. Thus, any interaction with the land-breeze front would most naturally be for the nocturnal warming to influence the land breeze rather than the other way around.

It seems that continuation of the sea breeze after sunset (under a favorable synoptic pressure pattern) had the effect of advecting a microscale temperature discontinuity (distinct from the sea-breeze front) that was established by rapid radiative cooling over land. Based on the wind speed and duration of the warming phase, temperature changes by advection only (at 2 m) would imply a 4°C coastal temperature gradient over a distance of about 1.7 km, which is possible if it is assumed that temperatures are uniformly around 15°C over the water (as observed at buoy 42007, 38 km to the southwest). Therefore, it is unclear what the relative roles of horizontal advection and turbulent mixing are for this case. What is clear is that dramatic changes of the surface layer regime occurred.

d. Case 4: Warming with cold-frontal passage and fog

On 8 November 2004 an interesting series of events occurred at Newton. For the previous several hours the surface layer was nearly saturated and most ASOS systems in the region reported fog or mist, with visibility as low as 0.4 km (0.25 mi). Winds were nearly calm, and skies were reported as clear above the fog. The KJAN sounding at 0000 UTC (1800 CST) indicated a strong surface inversion (5°C) already present in the region below about 100 m above ground level.
(AGL), with a moist layer (from 3 to 10 g kg\(^{-1}\)) below a weaker inversion at 1200 m AGL. Weak winds within the surface inversion changed to westerly winds of 6 m s\(^{-1}\) above. Around 0300 CST a dry cold front passed the Newton mesonet site (Fig. 14), with the sudden change to relatively strong northerly winds increasing the mixing sufficiently to result in a dramatic warming of the surface layer (Fig. 15). With a relatively well developed fog layer, the dewpoint initially increased until (presumably) the fog layer was completely dissipated. The increase in moisture content corresponds to a specific humidity increase of about 1.6 g kg\(^{-1}\). The temperature did not cool again sufficiently to reach the antecedent temperature before sunrise. Warming began first at 10 m, and there was actually a slight strengthening of the inversion as the fog dissipated. Most observing sites in the region also experienced an initial temperature rise following the frontal passage, with skies remaining clear. Although the inversion dissipation complicates identification of the front in the mesoscale analysis, a relatively clear pattern emerges when dewpoint depression is plotted (Fig. 16). Similarities

**FIG. 10.** As in Fig. 7 but at Calhoun City on 3 Mar 2006.
exist between this case and the cold-front-induced hydraulic jump described by Shapiro (1984) and Shapiro et al. (1985), but the practical impact on local conditions is somewhat different because of the strong impact on visibility as the fog was dissipated. The potential relationship to the evolution of density current wave patterns such as noted by Knupp (2006) or Rauber et al. (2001) is not clear.

Another interesting aspect of this day is that in the following evening, there were three moderately strong warming events within the cold air mass. Such nights with multiple short events have been noted on several other occasions, as the surface layer apparently struggles to achieve a stable equilibrium or possibly is forced by waves propagating on the inversion surface. The relatively short lived nature of such warming events is also more similar to cases that have typically been reported in the literature.

5. Statistics and classification

The limited climatological statistics available so far show that sudden nocturnal warming events can occur at any time of the year in Mississippi but seem to be slightly more common during dry summer conditions when the area is dominated by the subtropical high pressure center. The challenge of why some nights have uninterrupted nocturnal cooling and others have periods of warming remains a difficult issue. As pointed out by McNider et al. (1995), very subtle differences in initial conditions may result in vastly different degrees of surface layer decoupling.

Among the four Mississippi Mesonet sites at which records were available during the period, there were notably fewer events observed at Pascagoula. This fact appears to be related to higher mean wind speeds and a relatively small diurnal temperature range at this coastal site. In particular, conditions that were in general most conducive to nocturnal warming at Newton (summer-time high pressure) tended to produce at Pascagoula a strong land-/sea-breeze circulation that inhibited both nocturnal cooling and nocturnal calming of the surface winds.

The cases discussed above can be considered as typifying general categories of nocturnal warming events, based primarily on temporal patterns of temperature and dewpoint. Whether such categories necessarily have a consistent relationship to specific physical mechanisms is beyond the scope of the present investigation. Use of a different moisture variable (such as specific humidity) in the definition would not change the nature of the correlation with temperature and would be less easily related to dew and fog formation. (The counterargument would be that specific humidity relates more clearly to heat and moisture budgets, which are not explicitly considered here.) Thus, one can define a type-1 event as having a negative correlation between temperature and dewpoint throughout the event and a rapid cooling at the end of the event. These events seem to be more commonly associated with the intermittent microscale bursting noted by other researchers, although a wider variety of time scales are noted in the Mississippi data. Type-2 events are typified by a strong anti-correlation between temperature and dewpoint during the actual warming but a more gradual cooling that may

![Time series of 2-m temperature (solid line) and dewpoint (dotted line) at Pascagoula on 27–28 Feb 2006. Shading indicates a warming event; \( t_0 \), \( t_1 \), and \( t_{\text{end}} \) are as in Fig. 4.](image-url)
begin almost immediately after the sudden warming. Although the type-2 event considered above was associated with a mesoscale disturbance within a stable air mass, this should not necessarily be considered typical. For type-3 events the only criterion is that the warming not be associated with decreasing dewpoint, so that a wide variety of possibilities may occur. (Such cases have been observed at all sites, not just near the coast.) One may also broadly consider a type-4 category in which nocturnal warming is clearly associated with passage of a known synoptic or mesoscale boundary (such as a cold front or convective outflow boundary). Although this basic classification scheme is convenient, it is recognized that a wide variety of events and physical mechanisms could be contained within each of the categories, particularly for types 3 and 4.

Because the proposed classification is based, for convenience, on readily observed or easily quantified characteristics, it is doubtful that there will be a universal relationship between category and physical mechanisms. In fact, the wide variety of behaviors seems to indicate that most nocturnal warming events in Mississippi are driven by a complex interaction of mechanisms that cannot easily be isolated. Instead, the proposed classification provides a basis for further climatological studies of such relationships, with the understanding that a combination of mechanisms can generally be expected to be contributing. It is also proposed that this simple scheme (which admittedly does not address heat bursts, katabatic flows, or other events not yet identified in Mississippi) is particularly amenable for developing a climatological description that is useful for examining impacts on visibility, minimum temperature forecasts, and so on. There could also be value in separating events that occur with humidity near saturation or according to wind direction.

The classification scheme used here can be related to the broader classification that Kurzeja et al. (1991) used for the nocturnal boundary layer in South Carolina. A classification system was also proposed by Van de Wiel et al. (2003) for the CASES-99 data. All of the cases considered here would be classified as “unsteady” nights by Kurzeja et al. and as within the intermittent regime of Van de Wiel et al. In a similar way, Mahrt et al. (1997) have suggested (based on seven nights of 10-m data from Kansas) that the nocturnal boundary layer can be generally divided into three distinct regimes: very strong stability, a transitional regime of enhanced heat fluxes, and weak stability. These regimes might be considered to correspond to the different time phases within the warming events discussed here (preonset strong inversion, rapid warming phase, and postwarming weak inversion)—in particular, for type-1 events. In LeMone
et al. (2003), data from CASES-97 indicated that a change in horizontal scales of temperature may be associated with such boundary layer regime changes. Note that none of these prior (more general) classification schemes for the nocturnal boundary layer were necessarily linked to predominance of specific physical mechanisms for initiation or maintenance of the regime types.

When normalized by the station period of record, Newton observed about 30% more events (75 per year) than the other three stations. Using a subjective categorization into small, moderate, and large magnitudes, about twice as many large events occur at Newton as at Pascagoula. Newton also had the most open wind exposure, possibly indicating that the intensity of the “bursting” winds may be more important for sudden warming than the influence of obstacles in reducing the preonset wind speed (to favor preonset surface cooling). Based on a three-category classification into type-1, -2, and -3 events (i.e., not considering connection to fronts), only about 10% were type 1 and 38% were type 2. The fact that 52% were type 3, the broadest category, is indicative of the wide range of events included in this category rather than
implying that there are necessarily a large number of events with evolution that is similar to that of case 3.

6. Theoretical context and implications

In the vast majority (about 90%) of cases examined, warming begins at 10 m before it does at 2 m. This is consistent with an increase of wind speed that originates above the surface layer. The cases noted by Acevedo and Fitzjarrald (2001) were attributed to reversal of the evening transition by increased turbulence, resulting in a recoupling of the surface layer. Comparison of stations within their microscale observing network indicated that this regime transition was primarily driven by local processes at individual stations. They also pointed out a change in sensible heat fluxes during the recoupled regime. This is consistent with the dramatic change in sensible and latent heat fluxes at Newton in case 1.

The question, of course, remains regarding the mechanism (local or otherwise) that triggers the increased wind speeds required to recouple the surface layer, erode the inversion, and mix down drier air. If there are indeed no nearby frontal systems or convective activity, then there is no obvious easily discernible mechanism. The presence of a weak convective cell on radar at 0145 UTC (1945 CST) about 50 km southwest of the Newton mesonet station in case 1 does suggest the possibility of a heat burst, although the lack of warming at other sites and weak wind speeds are not consistent with documented heat bursts. It is conceivable that in some cases gravity waves [such as those documented from CASES-99 by Sun et al. (2002, 2004)] or indirect forcing from weak upper tropospheric shortwave troughs (Uccellini and Koch 1987) may serve this role. Interaction between the near-surface flow and regional forest cover could itself tend to generate more common (though weak) gravity waves (Lee 1997). However, note that the density current events from CASES-99 discussed in detail by Sun et al. (2002) tended to be associated with near-surface cooling rather than with warming. Conditions under which the nocturnal boundary layer is particularly prone to episodes of bursting between relatively laminar and turbulent regimes have been discussed by ReVelle (1993) and are generally consistent with at least the type-1 and -2 cases discussed above. However, the cases considered by ReVelle were of relatively short duration. In the study of Acevedo and Fitzjarrald (2001), no attempt was made to provide a causal explanation for the sudden increase of turbulence that was observed. Note that the prototype cases discussed here resulted in warming that persisted over much longer periods of time than were noted by most

![Surface synoptic chart for 0600 CST 8 Nov 2004.](image)
previous investigations, although not all cases observed were so persistent.

Triggering of warming in case 4 by passage of a cold front is not altogether surprising but may have consequences that have yet to be fully explored for the relatively flat terrain of the region. As discussed by Saucier (1955, section 9.18), when a cold front passes over a stably stratified valley in mountainous terrain, the cold air mass behind the front may be less dense and may remain decoupled above the surface layer. In such cases, mixing is inhibited and the impact of the front at the surface is relatively gradual. However, in the 2004 frontal case at Newton there was negligible topography and substantial mixing following the frontal passage, so that the temperature initially warmed rapidly. The potential for a temperature increase following a cold front, due to turbulent mixing, has been considered for a shallow valley by Sanders and Kessler (1999). It is not clear whether the case of warming following passage of a nocturnal cold front in nearly flat terrain is primarily an artifact of location of observing sites in clearings within a heterogeneous forested landscape (thus giving some dynamical resemblance to a depression; Gustavsson et al. 1998) or if the inversion and warming have a vertical scale significantly larger than the forest canopy height.

With regard to cloud radiative forcing and condensation, neither seems to contribute significantly in the cases presented. Although no observations of sky cover are available from the mesonet stations, reports of clear conditions at nearby ASOS sites make cloud radiative forcing an unlikely mechanism (especially because it would be most effective with low base clouds). It has also been noted that temperature rises are normally observed at 10 m first. If cloud radiative feedback were a dominant mechanism, it might be expected that the warming would first be seen closest to the surface or would be nearly simultaneous. In addition, such a process would not be expected to be associated with increased wind speeds. In a similar way, latent heating by condensation of fog does not seem to contribute because only case 4 showed evidence of fog, and in that case the fog was already established long before the warming event. The potential role of latent heat from dewfall in the nocturnal heat budget has been discussed recently by Whiteman et al. (2007), although it appears more likely merely to reduce the rate of cooling rather than to contribute to an actual warming.

7. Summary

Four cases of rapid nocturnal warming have been discussed using data from the Mississippi Mesonet, illustrating some of the variety of scenarios observed. Although all events are characterized by a rapid warming (over less than 1 h), the return to antecedent temperature and the evolution of dewpoint vary among events. In general, there is good correspondence between warming and increased wind speed. There are also dramatic decreases in the strength of the temperature inversion between 2 and 10 m. With the exception of the case involving a cold front, there was no significant change in wind direction associated with the warming events. Although evidence points primarily to warming by increased turbulent mixing.

**Fig. 15.** Time series of 2-m temperature (solid line) and dewpoint (dotted line) at Newton on 8 Nov 2004. Shading indicates a warming event; \( t_0 \) and \( t_1 \) are as in Fig. 4; \( t_{end} \) corresponds to sunrise.
following an increase in wind speed, the mechanisms forcing the wind speed variations are not always clear. While it is known that some events are forced by synoptic and mesoscale fronts, there are many cases in which weak gravity waves propagating on the inversion surface would be a plausible mechanism. None of the cases exhibited characteristics typical of heat bursts, and cloud radiative forcing did not appear to be a significant factor. Although many events seem to be driven by mechanisms local to an individual station, there were also cases that were observed at multiple observing sites.

Using the four cases discussed here, a simple classification of warming events is proposed. Using data with sufficient temporal resolution, nonconvective and non-katabatic warming events can routinely be classified based on the pattern of evolution of the temperature and dewpoint (or other moisture variable) and relation to density discontinuities such as fronts. A preliminary statistical analysis indicates some variation in frequency of different types of nocturnal warming events among observing sites. Using this information, the influence of synoptic and geographic factors on warming events can be tentatively discussed. It is suggested that sudden nocturnal warming events may be of significance to such applications as pollutant fumigation, daily minimum temperature forecasts, and mesoscale/microscale variations of visibility.

Acknowledgments. The author was funded by NOAA Cooperative Agreement NA17AE1623. Helpful discussions with the staff of the National Weather Service Forecast Office in Jackson are acknowledged, as well as several useful suggestions of the reviewers. The Mississippi Mesonet Steering Committee and various other
partners have been crucial to the development of the Mississippi Mesonet. Flux data at Newton were obtained through the courtesy of Heping Liu and Yu Zhang of Jackson State University.

REFERENCES


