Factors Affecting Surface Wind Speeds in Gravity Waves and Wake Lows

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ABSTRACT

Ducted gravity waves and wake lows have been associated with numerous documented cases of “severe” winds (>25 m s⁻¹) and wind damage. These winds are associated with the pressure perturbations and transient mesoscale pressure gradients occurring in many gravity waves and wake lows. However, not all wake lows and gravity waves produce significant winds nor wind damage. In this paper, the factors that affect the surface winds produced by ducted gravity waves and wake lows are reviewed and examined. It is shown theoretically that the factors most conducive to high surface winds include a large-amplitude pressure disturbance, a slow intrinsic speed of propagation, and an ambient wind with the same sign as the pressure perturbation (i.e., a headwind for a pressure trough). Multiple case studies are presented, contrasting gravity waves and wake lows with varying amplitudes, intrinsic speeds, and background winds. In some cases high winds occurred, while in others they did not. In each case, the factor(s) responsible for significant winds, or the lack thereof, are discussed. It is hoped that operational forecasters will be able to, in some cases, compute these factors in real time, to ascertain in more detail the threat of damaging wind from an approaching ducted gravity wave or wake low.

1. Introduction

Numerous cases of high winds, sometimes causing wind damage, associated with either gravity waves or “wake lows,” have been discussed in the literature (e.g., Bosart and Seimon 1988; Bauck 1992; Koch and O’Handley 1997; Bradshaw et al. 1999; Gaffin 1999; Koch and Saleeby 2001). Loehrer and Johnson (1995) observed three cases of wind gusts in excess of 25 m s⁻¹ [considered “severe” by the National Weather Service (NWS)] in wake lows associated with mesoscale convective systems (MCSs) during the 1985 Oklahoma Preliminary Regional Experiment for Storm-Scale Operational and Research Meteorology (PRE-STORM). The surface winds in wake lows and gravity waves (unless otherwise noted, we are referring to ducted gravity waves in this study) are associated with the transient but large mesoscale pressure gradients occurring in many of these features. Table 1 lists some examples of gravity wave–wake low events that produced significant observed surface winds and/or wind damage.

However, as demonstrated by Loehrer and Johnson (1995), many wake lows do not produce significant winds. The same may be said of ducted gravity waves. Gravity waves are referred to as ubiquitous mesoscale features of the atmosphere by Jin et al. (1996), and wave ducts are common. Clearly, damaging winds are not ubiquitous. It is the purpose of this paper to review the factors affecting the surface winds produced by gravity waves and wake lows. It will be shown that these factors include the amplitude of the pressure disturbance, its intrinsic speed of propagation (relative to the mean ambient wind), and the ambient wind itself.

Section 2 will include a review of gravity waves and wake lows, and discusses briefly the parallels between them and the fact that some authors indicate a wake low may be a form of a gravity wave phenomenon. It is beyond the scope of this paper to determine this, and it is also unnecessary for this study because both ducted gravity waves and wake lows produce surface winds in the same way, through vertical motion, subsequent surface pressure perturbations, and the highly ageostrophic wind response to these perturbations. The winds of either phenomenon may be described by the impedance relation, as will be shown. Section 3 uses theory to analyze the winds produced by these features. Section 4 contains several case study comparisons, illustrating how multiple
factors may contribute to damaging winds in a gravity wave or wake low, or the lack of damaging winds. Section 5 contains conclusions.

2. Review of gravity waves and wake lows

a. Gravity waves

Gravity waves are simply waves for which the restoring force is buoyancy (e.g., Lindzen 1990). As Hooke (1986) points out, much like the atmosphere transmits pressure disequilibria through sound waves, it transmits density disequilibria through gravity waves. Internal gravity waves may be initiated in the atmosphere by many different phenomena, including topographic forcing, vertical wind shear, geostrophic adjustment processes, and convection (Koch and O’Handley 1997). A type of gravity wave phenomenon that may affect the sensible weather at the surface, and is being examined herein, is the ducted gravity wave (e.g., Lindzen and Tung 1976).

A ducted wave is essentially the superposition of two internal gravity waves, one propagating upward and one propagating downward, often trapped in a “duct” between the surface and some reflecting level in the atmosphere. The reflecting level is associated with some combination of a decrease in static stability and/or curvature in the wave-relative wind profile, such as that associated with a low-level jet (Scorer 1949; Nappo 2002; Gill 1982). Lindzen and Tung (1976) show that when the depth of the duct is \( \frac{1}{4} \) of a vertical wavelength of the internal waves, the upward- and downward-propagating waves constructively interfere.

The basic kinematics of a ducted gravity wave are illustrated by Coleman (2008) (see Fig. 1). Note that subsidence warming ahead of the wave trough produces a negative surface pressure perturbation in the trough, while adiabatic cooling produces a positive \( p’ \) in the wave ridge. In an idealized ducted gravity wave, the perturbation winds and pressure at the surface are correlated

<table>
<thead>
<tr>
<th>Date</th>
<th>Location</th>
<th>Amplitude (( p’ ), hPa)</th>
<th>Max observed wind gust (m s(^{-1}))</th>
</tr>
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<tr>
<td>27 Feb 1984</td>
<td>Raleigh, NC</td>
<td>12</td>
<td>29</td>
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<tr>
<td>24 June 1985</td>
<td>Wakeeney, KS</td>
<td>N/A</td>
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<tr>
<td>7 Sep 1990</td>
<td>Olympia, WA</td>
<td>4</td>
<td>17</td>
</tr>
<tr>
<td>11 Apr 1995</td>
<td>Little Rock, AR</td>
<td>9</td>
<td>27.5</td>
</tr>
<tr>
<td>28 Apr 1996</td>
<td>St. Louis, MO</td>
<td>5</td>
<td>31</td>
</tr>
<tr>
<td>22 Feb 1998</td>
<td>Birmingham, AL</td>
<td>8</td>
<td>22.7</td>
</tr>
<tr>
<td>12 Apr 2005</td>
<td>Birmingham, AL</td>
<td>3.7</td>
<td>13</td>
</tr>
<tr>
<td>12 Apr 2005</td>
<td>Montgomery, AL</td>
<td>4.7</td>
<td>20.8</td>
</tr>
<tr>
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<td>6</td>
<td>23.5</td>
</tr>
<tr>
<td>7 Mar 2008</td>
<td>Calera, AL</td>
<td>4</td>
<td>19.4</td>
</tr>
</tbody>
</table>

\( a \) Bosart and Seimon (1988);  
\( b \) Loehr and Johnson (1995);  
\( c \) Bauck (1992);  
\( d \) Gaffin (1999);  
\( e \) Bradshaw et al. (1999).

Fig. 1. Airflow vectors in the \( x-z \) plane associated with a ducted gravity wave (horizontal wavelength of 50 km, duct depth of 2 km). The solid curve represents an isentrope, and the capital letters H and L indicate the positions of the maximum and minimum surface pressure perturbations, respectively. Large hollow arrows represent vertical motions, and large gray arrows represent maximum wind perturbations near the surface (adapted from Coleman 2008).
Forecasting gravity wave formation is quite difficult. Of the four processes that commonly initiate internal gravity waves listed by Koch and O’Handley (1997), vertical wind shear and geostrophic adjustment are probably the most common causes of ducted gravity waves producing significant surface wind away from mountainous regions. Vertical shear instability may produce Kelvin–Helmholtz waves in layers where the Richardson number Ri < 0.25 (e.g., Batchelor 1967; Miles 1961), and some authors (e.g., Scinocca and Ford 2000) have shown that this process may lead to internal gravity waves that propagate away from the shear layer. Koch and O’Handley (1997) summarize the favorable synoptic situation for the generation of internal gravity waves due to geostrophic adjustment. The waves must also typically be ducted in order to produce significant effects at the surface, and the duct factor (Koch and O’Handley 1997) is readily computed, and is automatically output by many NWS Advanced Weather Interactive Processing System (AWIPS) workstations.

b. Wake lows

Wake lows are local pressure minima located at or near the back edge of the trailing stratiform precipitation region of an MCS (e.g., Johnson and Hamilton 1988; Locher and Johnson 1995). They are in the “wake” of the MCS, hence the term wake low. They are often associated with a sharp gradient in radar reflectivity (e.g., Stumpf et al. 1991; Johnson et al. 1989). Wake lows (also known as wake depressions) were discussed and/or observed early on by Brunk (1953), Fujita (1955, 1959, 1963), and Williams (1953). The basic dynamics of a wake low are shown in Fig. 2. The wake low is associated with subsidence and the resultant warming (e.g., Johnson and Hamilton 1988; Stumpf et al. 1991). Johnson and Hamilton (1988) and Stumpf et al. (1991) propose that the subsidence is associated with a descending rear-inflow jet. Stumpf et al. (1991) estimated, using a dual-Doppler analysis of an MCS with an associated wake low, that the relatively dry air in the rear-inflow jet (dry air was shown near 5 km MSL in a proximity sounding), cooled rapidly through evaporation upon encountering the stratiform precipitation. This air then descended rapidly due to negative buoyancy. In reference to wake lows, Haertel and Johnson (2000) and Miller and Betts (1977) also discuss low-level cooling in the stratiform precipitation region, while Gallus (1996) discusses cloud microphysical processes. However, Schmidt and Cotton (1990) propose a connection between gravity waves and the subsidence in wake lows, Haertel and Johnson (2000) state that a wake low may be described as a gravity wave phenomenon, and Pandya and Durran (1996) propose that the mesoscale circulation associated with a squall line may be caused by gravity waves. Therefore, it is possible that wake lows are indeed similar to gravity waves. For the purposes of this study, a more extensive discussion of the mechanisms that produce the subsidence and associated surface pressure perturbations in wake lows is not vital, nor is a discussion of the relationship between wake lows and gravity waves, since this study is concerned with the low-level response of the wind to the perturbation itself. The perturbation wind and pressure in wake lows also follow the impedance relation.

3. Winds associated with these disturbances

a. Wind perturbations

The winds coincident with the passage of wake lows and gravity waves are caused by the transient pressure gradients associated with these disturbances. We will approximate the disturbance in pressure as sinusoidal in x (the direction of propagation of the disturbance) and t, which is fairly accurate for ducted gravity waves and most wake lows (e.g., Vescio and Johnson 1992). (Even if the disturbance is not sinusoidal in nature, as long as it is periodic in x and t, the following is still valid.) We will also assume it is form preserving (i.e., constant amplitude). Therefore, the linear impedance relationship (e.g., Gossard and Munk 1954; Gossard and Hooke 1975) applies, and may be written as (e.g., adapted from Koch and Golus 1988)

\[ u' = \frac{p'}{\rho(c - U)}. \]  

where \( u' \) is the wind perturbation, \( p' \) is the pressure perturbation, \( \rho \) is the density, \( c \) is the ground-relative speed of the disturbance, \( U \) is the component of the background wind in the direction of propagation of the disturbance, and \( c - U \) is the intrinsic propagation speed (relative to the mean flow). By convention, the background wind is positive (\( U > 0 \)) when it is in the same direction as the propagation of the disturbance, and negative (\( U < 0 \)) when it is in the opposite direction. We are examining winds at the surface, so to a first approximation, \( w = 0 \). The impedance relation is true independent of the wavelength or spatial scale of the disturbance. A disturbance with a given amplitude and intrinsic speed will have larger pressure gradient accelerations at smaller wavelengths, but the accelerations will last over a shorter duration.

Equation (1) implies several important points regarding the surface wind perturbations in a gravity wave or wake low. First, we note that \( p' \) and \( u' \) are correlated, as...
mentioned in section 2a. This is true as long as the disturbance is not experiencing a tailwind that is also larger in magnitude than the wave’s ground-relative speed \( c \) (i.e., \( U > c \)); this would reverse the correlation, but significant disturbances with this property are rarely observed at the surface. Therefore, larger-magnitude pressure perturbations produce larger-magnitude wind perturbations, and the perturbations in a pressure ridge (trough) are positive (negative). Second, slower-moving disturbances (i.e., small value of \( c - U \)) cause larger-magnitude wind perturbations. This makes physical sense, as parcel residence times are longer in slow-moving disturbances than in faster-moving ones. It may be shown that Eq. (1) is true for any linear, form-preserving, periodic disturbance in pressure. Equation (1) is not exactly valid if the disturbance is not form preserving, such as in a case where the amplitude is increasing with time. However, Eq. (1) still provides a good approximation of the amplitude of the wind perturbation \( u' \) even in growing–decaying disturbances; the correlation is not exact, as the peak \( u' \) leads the peak \( p' \) in phase (see Fig. 6 in Vescio and Johnson 1992).

A simple numerical model was used to apply the linear, inviscid, irrotational horizontal equation of motion [Eq. (2)], to a form-preserving pressure disturbance that is periodic in \( x \) and \( t \), as follows:
where $U$ is the component of the background wind parallel to the motion of the disturbance. The following transformation, which is also valid for form-preserving disturbances, was used:

$$\frac{\partial u'}{\partial t} = -\frac{1}{\rho} \frac{\partial p'}{\partial x} - U \frac{\partial u'}{\partial x}.$$  \hspace{1cm} (2)

Resulting winds simulated in this numerical model, applied to three waveforms of amplitude 1 hPa with no background wind (assuming a density of 1.2 kg m$^{-3}$), are shown in Fig. 3. Note that, for the sinusoidal, sawtooth, and hyperbolic secant pressure waves, pressure and wind perturbations are correlated and are consistent with the impedance relation. The hyperbolic secant waveform is similar to that derived for a shallow water solitary wave (e.g., Boussinesq 1871; Rayleigh 1876).

b. Total wind speed

An important point to consider is that the total component of the wind parallel to the wave motion ($u$) is given by the sum of the background and perturbation winds at the surface (i.e., $u = u' + U$). This, in addition to the component of the surface wind normal to wave motion ($v$), is what will be observed. In many cases of high surface winds caused by wake lows and gravity waves, the component of the surface wind normal to the wave motion is not significant when compared to $u'$, unless it is in excess of 10 m s$^{-1}$. Since $p'$ and $u'$ are correlated, the largest surface wind speeds are achieved for a given disturbance when the sign of the background wind is the same as the sign of the pressure perturbation. This makes physical sense. If $U > 0$, then the highest wind speeds will occur where $u' > 0$ also, that is, in the pressure ridge ($p' > 0$). If $U < 0$, then the highest wind speeds will occur in the pressure trough, where $u'$ and $p'$ are both negative. In summary, for pressure troughs, the magnitude of the surface wind is proportional to the amplitude, inversely proportional to the intrinsic speed ($c - U$), and increases as $U$ decreases, that is, as the trough is moving into a stronger headwind ($U < 0$) or a weaker tailwind ($U > 0$).

The maximum wind speeds in pressure troughs, calculated using the impedance relation, as their amplitude ($p'$), intrinsic speed ($c - U$), and background wind ($U$) vary, are shown graphically in Figs. 4–6. In Fig. 4, a background wind of $U = 0$ was assumed. Figure 4 shows that, for a given background wind, the maximum surface wind speed increases with amplitude $p'$ and decreases with increasing intrinsic speed ($c - U$). Figure 5 shows that, at a given amplitude $p'$ (in this figure, $p' = 1$ hPa), surface wind increases as the headwind increases, and decreases with intrinsic speed. Figure 6 shows the necessary $p'$ required to produce 25 m s$^{-1}$ (50 kt, defined as severe by the National Weather Service), as the intrinsic speed and the background wind vary. As expected, larger pressure perturbations are required to produce severe winds for fast-moving wave troughs, and for troughs with tailwinds ($U > 0$) as opposed to headwinds ($U < 0$). In summary, when friction is neglected, large-amplitude, slow-moving disturbances, whose pressure perturbation is of the same sign as the disturbance-relative background wind, produce the largest surface wind speeds.

It should be pointed out that the calculation of $p'$ and $U$ is not always straightforward. For example, when determining $p'$ in an environment where there is also a significant synoptic pressure tendency, the perturbation is the deviation of the pressure from its background trend. The assumption here is that the large-scale wind field remains adjusted to the longer-term pressure tendency, but the winds in waves and wake lows are due to the transient, large perturbations from the background trend. This is more clearly illustrated in Fig. 7. Regarding the background wind $U$, one must assess the low-level background flow away from the influence of a ducted gravity wave, or from a wake low and the MCS associated with it. For example, as shown in Fig. 2b, an MCS often contains a mesoscale perturbation high pressure area [termed a mesohigh by Johnson and Hamilton (1988)], near the convective line, ahead of the wake low. This positive $p'$ will tend to locally accelerate winds in the direction of MCS motion. Therefore, the winds within the MCS are not representative of the background wind $U$.

4. Case studies

Each of the following pairs of case studies illustrates how a given variable (background winds, intrinsic speed, or amplitude of the pressure disturbance) may affect the surface winds. In each pair of cases, one disturbance produced significant winds or wind perturbations, while the other did not.

a. Background winds

The following two cases illustrate the difference the ambient winds can make in determining whether or not a disturbance with a given wind perturbation will produce high observed winds. Two gravity waves–wake lows are shown, both having similar wind perturbations. However, one primarily produced only a shift in wind direction without significant wind speeds, while the other
FIG. 3. Here, $p'$ (solid curve) and $u'$ (dashed curve) are simulated by a linear model (see text) for (a) sine waves, (b) sawtooth waves, and (c) tanh waves (similar to solitary waves; see text). In each case, the horizontal wavelength is 30 km and the wave speed is 15 m s$^{-1}$, implying a wave period of 2000 s.
produced widespread wind damage. Both appeared to be associated with descending rear-inflow jets and were located near the back edges of MCSs.

1) 10 MAY 2006

On 10 May 2006, a wake low–gravity wave propagated across northern Alabama, near the back edge of an MCS. The wake low followed a mesoscale high pressure area associated with the MCS. Its amplitude was 3.5 hPa, and this amplitude was fairly consistent at multiple surface stations over which the disturbance passed, from Huntsville to Gadsden, Alabama, which is located about 100 km to the southeast of Huntsville. Therefore, the disturbance may be approximated as two-dimensional. Radar and surface data indicate that the disturbance was propagating from 260° at 24 m s⁻¹. Background winds at low levels (ahead of the MCS and the wake low) were generally light and southerly, at speeds less than 5 m s⁻¹. In this case, the component of the background wind in the direction of disturbance propagation was very nearly $U = 0$, since the background winds are also normal to the motion of the wake low. Therefore, $p' = -3.5$ hPa and $c - U = 24$ m s⁻¹. The density was calculated to be $\rho = 1.18$ kg m⁻³, based on the pressure (not adjusted to mean sea level) and temperature at Huntsville. The impedance relation would therefore predict a maximum wind perturbation of $u' = -12.4$ m s⁻¹, and since $U = 0$, the maximum wind speed should also be near 12 m s⁻¹.

Time series of the pressure, the component of the wind in the direction of disturbance motion $u$, and the total wind speed at the Huntsville International Airport (HSV) Automated Surface Observing System (ASOS) station are plotted in Fig. 8. Notice that the MCS was associated with a 3-hPa pressure ridge (between 1700 and 1815 UTC), associated with a small positive wind perturbation as winds gained a more westerly component. Then, a rapid pressure fall of 6.5 hPa occurred in only 19 min, between 1818 and 1837 UTC, as the wave–wake low passed. The pressure rise after 1930 UTC was associated with another MCS. There is excellent correlation between $p$ and $u$, consistent with the impedance relation, and the largest-magnitude wind perturbation occurred at 1837 UTC, within 1 min of the end of the rapid pressure drop. The component of the maximum wind perturbation in the direction of the disturbance motion, $u' = -12.5$ m s⁻¹, is also very close to that predicted by the impedance relation. The maximum wind speed (Fig. 8c) at the ground is from the east (95°) at 13.3 m s⁻¹; this wind is in roughly the opposite direction of the direction of propagation, and within 1 m s⁻¹ of the magnitude of $u'$. This is due to the lack of background wind in the wave-parallel direction ($U = 0$).
2) 28 APRIL 1996

A wake low–gravity wave propagated across parts of Missouri on 28 April 1996, producing wind gusts in excess of 25 m s$^{-1}$ (50 kt) at a number of locations (Gaffin 1999). Wind gusts as high as 31 m s$^{-1}$ were reported in the St. Louis area (Gaffin 1999). This disturbance was also at the rear of an MCS. The pressure perturbation associated with this disturbance was $-5$ hPa at St. Louis (STL; see Fig. 9), and the surface wind perturbation at STL was approximately $u' = -15$ m s$^{-1}$, only about 2.5 m s$^{-1}$ higher than in the 10 May 2006 case. However, surface observations from STL before the disturbance passed indicate that the component of the background wind in the direction of disturbance motion was $U = -10$ m s$^{-1}$. Therefore, the total component of the wind in the direction of the wave motion in this case was $u = -25$ m s$^{-1}$ (50 kt).

Vertical profiles of the component of the wind in the direction of disturbance motion, associated with both cases, were produced using smoothed cross sections (normal to the disturbance) of velocity data from Doppler radars (Fig. 10). Background winds were also determined using Doppler velocity data well ahead of the disturbance. These profiles allow comparison of the winds in the direction of the wave motion ($u$), and perturbation winds ($u'$), at various altitudes. Note that on 10 May 2006, when the background wind was near zero at the surface and in the direction of propagation of the disturbance ($U > 0$) between 0 and 500 m AGL, but the wind perturbation was in the opposite direction ($u' < 0$), the two partially canceled each other out (Fig. 10a), and the total low-level wind was less than 15 m s$^{-1}$ (Fig. 10b). However, on 28 April 1996, the background wind and the wind perturbation were both in the same direction (both in the opposite direction of the disturbance motion, so $U < 0$ and $u' < 0$); therefore, they combined to produce much stronger low-level winds near 35 m s$^{-1}$, fairly close to the observed maximum winds of 31 m s$^{-1}$ in the St. Louis area (Gaffin 1999). In both cases, the wind perturbation in the pressure trough was near $u'' = -15$ m s$^{-1}$, but the direction of the background wind made the difference between an insignificant wind shift on 10 May 2006 and damaging winds on 28 April 1996.

b. Intrinsic speed of disturbance

Here, two disturbances with similar amplitudes, but different intrinsic speeds, are examined. It will be shown that the disturbance moving slower relative to the mean wind produced a larger wind perturbation. The first
FIG. 8. One-minute surface observations at HSV on 10 May 2006 of (a) pressure (hPa), (b) the component of the wind in the direction of the wake low propagation (m s$^{-1}$), and (c) wind speed (m s$^{-1}$).
disturbance being examined is the 10 May 2006 gravity wave–wake low in north Alabama, discussed in section 4a(1). As discussed, in that event, $p' = -3.5$ hPa, the intrinsic speed of the disturbance was $c - U = 24$ m s$^{-1}$, and the maximum wind perturbation in the direction of disturbance motion was $u' = -12.5$ m s$^{-1}$.

This event is similar in magnitude to a gravity wave that passed through parts of Missouri on 1 December 2006. An excellent wave duct was in place, with the 0600 UTC North American Mesoscale (NAM) model sounding at St. Louis (Fig. 11) showing a deep layer of stable air up through 750 hPa, with less stable air above that, consistent with the ducting theory in Lindzen and Tung (1976). The wave was propagating through an area of precipitation, and was associated with enhanced radar reflectivity near the wave ridge, and decreased reflectivity near the wave trough (Fig. 12). The wave produced a surface pressure ridge at STL of $p' = 3$ hPa (Fig. 13), very similar to the amplitude in the 10 May 2006 case. The wave was propagating at a ground-relative speed of 25 m s$^{-1}$, but the average of the component of the background wind in the direction of wave motion was $-5.9$ m s$^{-1}$.

Therefore, the intrinsic phase speed $c - U = 30.9$ m s$^{-1}$, or 29% faster than the 10 May 2006 case. As one may infer from Fig. 13, the maximum wind perturbation was $u' = 7.6$ m s$^{-1}$. The wind perturbation was positive since it was associated with the wave ridge, consistent with the impedance relation. Therefore, the magnitude of the maximum wind perturbation was 40% less in the 1 December 2006 case than in the 10 May 2006 case. This is likely related to the fact that the 1 December 2006 wave was propagating at a significantly faster intrinsic speed; this implies smaller wind perturbations due to shorter parcel residence times, as discussed in section 3a.

c. Amplitude of the pressure disturbance

The following two cases illustrate the large effect that the magnitude of the pressure perturbation has on the wind perturbation and the maximum wind speed.

1) 20 DECEMBER 2007

On the afternoon of 20 December 2007, a fairly intense wake low–gravity wave propagated across much of northeast Mississippi and northern Alabama, at the rear edge of a large area of primarily stratiform rain to the north of an MCS (Fig. 14). One-hour surface pressure falls of 4–8 hPa and wind gusts $>15$ m s$^{-1}$ were common. There were numerous reports of wind damage and power outages across 11 Alabama and 3 Georgia counties. Alabama Power Company reported that 18 000 customers lost power in the Birmingham area (Elliott 2007). The damage occurred over a fairly long time period, with reports beginning in northwest Alabama as early as 1830 UTC 20 December, and lasting until 0225 UTC 21 December (NCDC 2007). The disturbance was propagating from 300° at approximately 13 m s$^{-1}$, relative to the ground, and at an intrinsic speed $c - U = 14$ m s$^{-1}$. The amplitude of the disturbance at Birmingham (BHM) was $p' = -5$ hPa. A disturbance with this amplitude and intrinsic propagation speed could produce ground-relative winds near 30 m s$^{-1}$, based on the impedance relation and background winds. The 1-min ASOS observations from BHM (Fig. 15) show wind gusts around 16 m s$^{-1}$, and higher gusts occurred at some, mainly elevated locations, including 23.5 m s$^{-1}$ at Inverness, Alabama, and 22.6 m s$^{-1}$ at Cullman, Alabama (Spann 2007).

2) 31 JANUARY 2008

Several apparent gravity waves propagated across northern Alabama during the early morning hours of 31 January 2008. A time–height section of potential temperature $\theta$, computed using temperature data from the University of Alabama in Huntsville’s (UAH) Microwave Profiling Radiometer (MPR), shows the vertical displacement of the isentropes associated with the waves around 0900 and 1055 UTC (see Fig. 16). Note also the much larger vertical gradients of potential temperature at low levels than at midlevels. A vertical profile of the Brunt–Väisälä frequency $N$ at 0700 UTC, using potential

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Fig. 9. Surface observations at STL on 28 Apr 1996 of (a) wind speed (kt) and (b) mean sea level pressure (hPa) (adapted from Gaffin 1999).
temperature data from the MPR (Fig. 16), shows very stable air at low levels, with an average value of $N$ below 1 km AGL of 0.024 s$^{-1}$, and an average $N$ of 0.021 s$^{-1}$ below 2 km AGL. However, above 2 km AGL, the air is less stable, with the average $N$ being 0.013 s$^{-1}$ between 2 and 4 km AGL. This decrease in stability aloft reduces the vertical loss of wave energy, and provides some wave ducting (e.g., Lindzen and Tung 1976). It is interesting to note the rapid ascent and descent of low-level isentropes in Fig. 16 (i.e., 282 K) within the ducted region, implying larger vertical motions, with the much slower changes in altitude of the isentropes at midlevels (i.e., 303 K).

Hydrostatically, the upward motion and cooling around 0900 and 1055 UTC caused positive surface $p'$, as indicated in observations from the UAH surface station (Fig. 17). Pressure and wind are measured at 5-s intervals, but 2-min average values of $u$ (component of the wind in the direction of disturbance motion) and wind speed are shown. The pressure–wind correlation implied by the impedance relationship is clear in this case also. Using radar and surface observations, it was determined that these waves were propagating at approximately 22 m s$^{-1}$, and, given the background wind at the surface, the intrinsic wave speed was $c = U = 22.5$ m s$^{-1}$. The pressure perturbations are rather small in this case. The largest pressure perturbation at UAH is associated with the wave around 1055 UTC, but even then, $p' = 1.4$ hPa, and $p' = 0.5$ hPa with the disturbance at 0900 UTC, based on a background pressure rise rate of 0.77 hPa h$^{-1}$ between 0800 and 1400 UTC. Given
a density of 1.27 kg m$^{-3}$, the impedance relationship would imply $u' = 1.8$ and 4.9 m s$^{-1}$ with the disturbances at 0900 and 1055 UTC, respectively. The actual maximum wind speed between 0800 and 1100 UTC was only 4.8 m s$^{-1}$, a speed not significant in terms of operational meteorology. Even if the $p' = 1.4$ hPa wave had been moving with a much slower intrinsic speed of only 10 m s$^{-1}$, a ratio using the impedance relationship implies it would have only produced winds around 10.8 m s$^{-1}$; if the wave ridge had experienced a tailwind of $U = 10$ m s$^{-1}$, the actual wind speeds would have remained below 15 m s$^{-1}$. Clearly, waves with small amplitudes like the ones on 31 January 2008 are not sufficient to produce damaging winds under most circumstances.

5. Discussion and conclusions

Numerous cases of gravity waves and wake lows producing significant surface winds and, in some cases, damage, have been reported both in the previous literature and in this paper. As shown, especially in the cases of the 28 April 1996 event in St. Louis, and the 20 December 2007 event in Birmingham, these waves–wake lows may be significant weather events, causing wind damage, power outages, and a threat to aviation. Since gravity waves and wake lows are common in the atmosphere, it is important to delineate the factors associated with waves–wake lows that have an effect on whether they produce significant surface winds or not. The impedance relationship provides an estimate of the wind perturbations as related to the amplitude and intrinsic speed of the disturbance, but background winds outside the pressure disturbance must also be considered, as they vectorally add to any perturbation winds associated with the wave–wake low.

Perhaps the most straightforward factor in determining the damaging wind potential of a gravity wave or wake low is its amplitude. The wind perturbation is directly proportional to the amplitude, so the pressure perturbations associated with a disturbance must be ascertained. Although prediction of the amplitude of a gravity wave or wake low is difficult, Coleman (2008), combining equations from Holton (1992) and Lighthill (1978), showed that the surface pressure perturbation due to the vertical motion in a disturbance is proportional to the static stability of the atmosphere and to the vertical displacement of air parcels. Therefore, gravity waves and wake lows have the potential to have larger amplitudes in environments characterized by static stability at low levels. Also, the amplitude of a wake low,
especially in view of the findings of Pandya and Durran (1996), may be related to the intensity of the thermal forcing in the MCS, represented by the intensity of the MCS. Dry air at midlevels may also enhance the possibility that the air in the rear-inflow jet of an MCS will subside rapidly upon encountering precipitation, producing a large-amplitude pressure perturbation (e.g., Stumpf et al. 1991).

Another important factor is the component of the background wind outside the influence of the gravity wave–wake low in the direction of the wave propagation $U$, specifically whether the disturbance is experiencing a tailwind ($U > 0$), or a headwind ($U < 0$). In general, the background winds will enhance the surface wind speeds associated with a pressure disturbance when the sign of the background wind is the same as the sign of the pressure perturbation, since the disturbance-parallel component of the wind experienced at the surface is given by $u = U + u'$. For wake lows and gravity waves with negative pressure perturbations ($u' < 0$), a headwind ($U < 0$) will enhance the surface wind speeds, and a tailwind ($U > 0$) will reduce the surface wind speeds, perhaps instead only causing a change in wind direction, as opposed to a significant wind speed. The opposite is true for positive pressure perturbations, such as waves of elevation. In these cases, $u' > 0$, and a tailwind ($U > 0$) enhances the surface wind speed.

A clear example of the importance of background wind is shown in section 4a. The 10 May 2006 Huntsville, Alabama, disturbance and the 28 April 1996 St. Louis, Missouri, disturbance both had negative wind perturbations with magnitudes of 12.5–15 m s$^{-1}$. However, on 10 May 2006 the background winds were small and positive, and the disturbance produced maximum surface winds of only 13.3 m s$^{-1}$. On 28 April 1996, there was a significant headwind ($U = -10$ m s$^{-1}$), and the result was surface winds in excess of 25 m s$^{-1}$ and wind damage.

Another component shown to affect the wind perturbations in a wake low or ducted gravity wave is the intrinsic speed of propagation relative to the background wind, $c - U$. In a slow-moving disturbance, parcels experience longer residence times in the transient pressure gradients associated with the disturbance and, therefore, accelerate to larger magnitudes of $u'$. In faster-moving disturbances, parcel residence times are smaller, not allowing as much time for acceleration and, therefore, smaller magnitudes of $u'$. In section 4b, two
disturbances with similar amplitudes of 3–3.5 hPa were examined. The one that was propagating at an intrinsic speed of 24 m s$^{-1}$ produced wind perturbations of 12.5 m s$^{-1}$, but the one that was moving faster relative to the background flow (\(c - U\)) = 31 m s$^{-1}$ only caused wind perturbations of 7.6 m s$^{-1}$. The intrinsic speed of a ducted gravity wave may be anticipated, to some degree, in advance (Lindzen and Tung 1976), and is proportional to the depth of the wave duct and the static stability. However, this is not the case for wake lows; the speed must be assessed once the disturbance develops.

In summary, large-amplitude disturbances (large magnitude of \(p'\)) with background winds in phase with the pressure perturbation (\(U < 0\), or headwinds, for \(p' < 0\); \(U > 0\), or tailwinds, for \(p' > 0\)) are the most likely to produce damaging surface winds. Disturbances moving slowly (relative to the background wind) with similar amplitudes and background winds are also more likely

![Birmingham, AL (BHM) ASOS](image)

**Fig. 15.** The 1-min surface observations of pressure (hPa, solid curve), wind speed (dashed curve, m s$^{-1}$) and wind gust (dots, m s$^{-1}$) on 20 Dec 2007. Wind gusts are only shown when they were more than 2.5 m s$^{-1}$ greater than the wind speed.

![UAH MPR Potential Temperature](image)

**Fig. 16.** (a) Time–height section of potential temperature on 31 Jan 2008, derived from UAH MPR measurements. Note the vertical displacements of the isentropes around 0900 and 1055 UTC. (b) Profile of \(N\) at 0700 UTC, based on MPR data shown in (a).
to produce wind damage. With the increasing availability of mesoscale observations of pressure, but fewer accurate observations of wind (due to shielding by trees and other issues), the above factors may help forecasters determine the threat of damaging winds associated with an approaching gravity wave or wake low, even if they are not able to ascertain the surface winds near the disturbance readily. It should be noted that, while the background winds may be fairly constant, the amplitude of the disturbance and its intrinsic speed may, in some cases, change rather quickly, changing the parameters in the impedance relation. So, forecasters must continue to assess the speed and the amplitude of a ducted gravity wave or wake low as it propagates through their forecast area, watching for any changes that may increase (or decrease) the likelihood of damaging winds.

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