A Study of Wind Profiler Data Collected Upstream during Windstorms in Boulder, Colorado

J. BRENT BOWER AND DALE R. DURRAN

Department of Meteorology, University of Utah, Salt Lake City, UT 84112

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ABSTRACT

Wind profiler data from Lay Creek, Colorado, along with stability data from the Lander and Grand Junction rawinsonde observations, were examined in an attempt to link various parameters in the upstream flow to the onset of strong downslope winds in Boulder. Some correlation was found between the occurrence of high surface winds at Boulder and the upstream wind direction, upper tropospheric wind shear and the vertical phase shift across the troposphere. However, these parameters alone were not able to distinguish between windstorm and nonwindstorm events. It is likely that the remaining ambiguity could be eliminated with information on the location and strength of inversions in the upstream flow.

1. Introduction

Downslope winds have been observed in the lee of mountains all over the world. In some areas these winds gust to speeds exceeding 50 m s^{-1}, producing damage to buildings and impairing human activities. In addition, strong downslope winds are usually associated with large amplitude mountain waves which may pose a hazard to aircraft. In order to better understand and forecast this potentially destructive phenomenon, Colson (1954) and Brinkmann (1974) have examined the characteristics of the airflow upstream of the Sierra Nevada and Rocky mountains prior to, and during, downslope wind events. While their studies have proved useful to forecasters and researchers alike, their conclusions serve only as a general guide, and in many situations the difference between windstorm and nonwindstorm events still can not be unambiguously determined.

A major obstacle to these earlier studies was the coarse temporal and spatial resolution of the data, which were obtained primarily from the synoptic-scale rawinsonde network. A unique opportunity to examine higher-resolution data occurred when a wind profiler was operated at Lay Creek, Colorado during the 1983/84 winter. (An extensive description of the wind profiler instrument is given by Balsley and Gage, 1982.) As shown in Fig. 1, Lay Creek is located on the windward side of the Colorado Front Range, and is ideally situated to sample the upstream air flow when windstorms are occurring in the vicinity of Boulder. The profiler at Lay Creek provided hourly measurements of the horizontal wind field at 44 levels between the heights of 1690 and 17 600 m AGL. However, unlike a rawinsonde, the profiler provided no information about atmospheric stability. Stability data were obtained from the 12-hourly soundings at Grand Junction and Lander.

The purpose of this investigation is to examine the wind profiler data to determine whether characteristic signatures appear in the upstream wind field prior to downslope windstorms. According to theories proposed to explain downslope winds, the nature of the airflow is almost exclusively dependent on the atmospheric windspeed and stability, and the shape of the topography. Since the topography is fixed, the likelihood that windstorms can be linked to characteristic changes in the wind field will depend on the relative importance of changes in the stability.

Klemp and Lilly (1975) have proposed that downslope windstorms are produced by large amplitude mountain waves generated by the optimal superposition of vertically propagating waves which have been partially reflected at locations where there is an abrupt change in the Scorer parameter,

$$l^2 = \frac{N^2}{U^2} - \frac{1}{U}\frac{d^2U}{dz^2}, \qquad (1)$$

where N is the Brunt–Väisälä frequency and U is the horizontal wind component normal to the ridge crest. The most important "tuning" criteria is that there be a half-wavelength phase shift between the ground and the tropopause. Thus, according to this theory, the development of downslope winds will be primarily dependent on the integral properties of N and U. Since the integral of N over the depth of the troposphere exhibits far less temporal variability than the integral of U, one might expect that the changes in the upstream flow which produce a windstorm might be dominated by changes in the wind field.

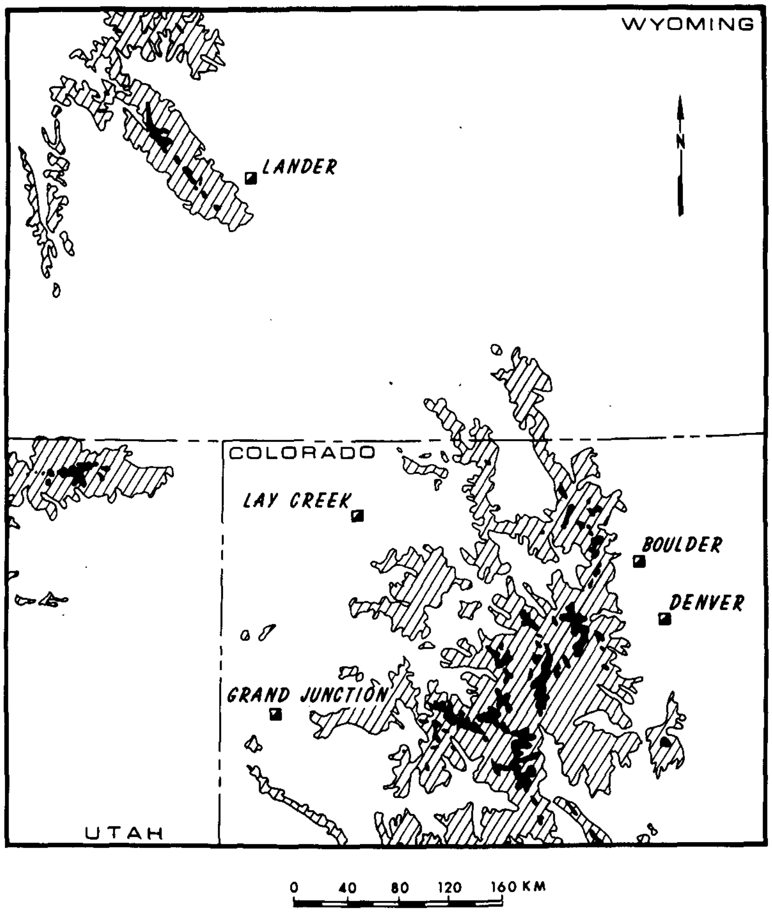

FIG. 1. Map showing the locations of Boulder, Grand Junction, Lander and Lay Creek. Terrain elevations over 9000 ft (2743 m) are shaded; elevations over 12 000 ft (3658 m) are solid black.

Although the theory of Klemp and Lilly provides theoretical support for the attempt to diagnose windstorms without access to fine resolution stability data, hydraulic theory is not so encouraging. Long (1953) has suggested that downslope winds are produced by a mechanism analogous to the transition from subcritical to supercritical flow which may occur in liquids passing over an obstacle. According to this theory, stability is primarily important in those regions where it is concentrated in near discontinuities (i.e., inversions). Although the temporal variation in the integral of N over the troposphere is small, the temporal variation in the height and strength of atmospheric inversions is considerable. Thus, according to Long's theory, the changes in the upstream flow which produce a windstorm are not likely to be dominated by changes in the wind field.

Peltier and Clark (1979) and Clark and Peltier (1977) have suggested a third theory, in which the occurrence of strong downslope winds is associated with the development of a breaking wave in the upper troposphere. If the upstream windspeed and stability do not vary with height, breaking waves will develop whenever the parameter Nh/U exceeds a threshold value of order 1. (The exact value depends on the terrain shape.) It may be possible to use an integrated form of this parameter to predict the development of breaking waves when the windspeed and stability vary with height. Then, since one is only using the integral values of N and U, it should again be possible to diagnose the development of windstorms without access to fine temporal resolution stability data.

2. Analysis based solely on profiler data

The basic design of this study is to identify windstorm events through surface observations and obtain the corresponding wind profiler and rawinsonde data for analysis. As a first step, the simple hypothesis was examined that windstorm events can be distinguished by signatures or patterns in the wind data alone. Measurements of the upstream airflow were collected from the wind profiler radar at Lay Creek, Colorado, which is approximately 250 km northwest of Boulder. Hourly averages of windspeed and direction were reported in two data sets with vertical resolutions of 290 and 870 m. The coarse and fine resolution data was analyzed separately, and also as a combined data set. The surface observations used to determine the windstorm periods were the anemometer traces from the NCAR tower on Table Mesa, on the west side of Boulder. The study consisted in analyzing data from seven cases. Two cases were moderately strong wind storms, four were weaker, and one nonwindstorm case was included for comparison. The nonwindstorm case occurred in a situation with strong westerly flow across the mountains, which appeared to favor the development of mountain waves and downslope winds.

The wind speed and direction, and the magnitude and direction of the wind shear vector, were examined to determine whether they exhibited any distinctive behavior associated with the onset of windstorms. The diagnosis of the winds was conducted using the tropospheric mean, and values at three separate altitudes: low-level (at 650 mb, near the mountain top), at the tropopause, and at the level of the fastest jet stream winds. The hourly variation in the tropopause height over Lay Creek was linearly interpolated in time and space between the heights observed in the Grand Junction and Lander soundings. This value was found to have good agreement with the level of maximum winds given by the profiler. Wind shear was calculated for the upper and lower troposphere as the difference between the tropopause and 500 mb winds, and the difference between the 500 mb and low-level winds, respectively. The surface windspeed, plotted in the following figures, is the maximum gust recorded during each 20-minute period by the NCAR anemometer.

The time dependent behavior of the speed of the east–west (cross mountain) wind component at the jet stream and 650 mb levels is compared with the surface wind gust data from Boulder in Figs. 2 and 3. A time lag of approximately three hours should be allowed for the time required for an air parcel sampled at Lay Creek to travel to Boulder. These figures do not suggest any simple relationship between changes in the upstream windspeed and the development of windstorms, nor is there a significant difference between the windspeed in the windstorm and nonwindstorm cases. The mean tropospheric windspeed and the wind data from other levels were also examined, but they failed to reveal any variations which appeared to correlate with the onset of downslope winds. Notice that strong jet stream winds are not required for the production of strong downslope winds in Boulder; however, during all the windy periods except the one centered at 00Z in case I, the cross-mountain component of the 650 mb wind exceeded 10 m s^{-1}. In every windstorm, the maximum wind gusts at the surface exceeded the mean 650 mb cross-mountain windspeed.

As with the windspeed, the wind direction did not show any distinct pattern associated with the onset of the windstorms; however, the direction of the upper tropospheric wind was almost always between 280° and 315° while strong downslope winds were blowing. The criterion that the strongest winds occur in Boulder when the upper level winds are slightly north of west has been previously noted by Brinkmann (1974). Since the mountains west of Boulder run essentially north–south, it is surprising that the preferred wind direction is north of due west. This may be due to the fact that the highest mountain peaks in the vicinity of Boulder are actually somewhat northwest of the town. It may also be due to climatological differences in the stability structure associated with west-southwesterly and west-northwesterly flow. In the case of a west-southwesterly upper tropospheric flow, Boulder is likely to be located downstream of the trough axis in a synoptic scale wave. The synoptic scale vertical motions in this portion of the wave are usually upward, and would, therefore, act to reduce the stability of the lower troposphere. In the case of a west-northwesterly flow, Boulder is likely to be upstream of the trough axis, a location in which there is often synoptic-scale descent which would act to increase the lower tropospheric stability.

The time dependent behavior of the wind direction

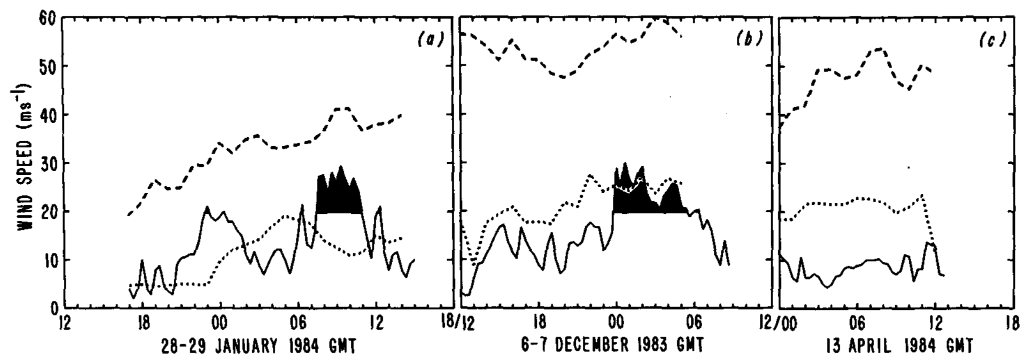

FIG. 2. East–west wind component at the 650 mb (dotted) and jet stream level (dashed), and surface wind gusts in Boulder (solid) for (a) case I, (b) case II and (c) case VII. The portion of the surface wind trace which exceeds 20 m s^{-1} is highlighted in black.

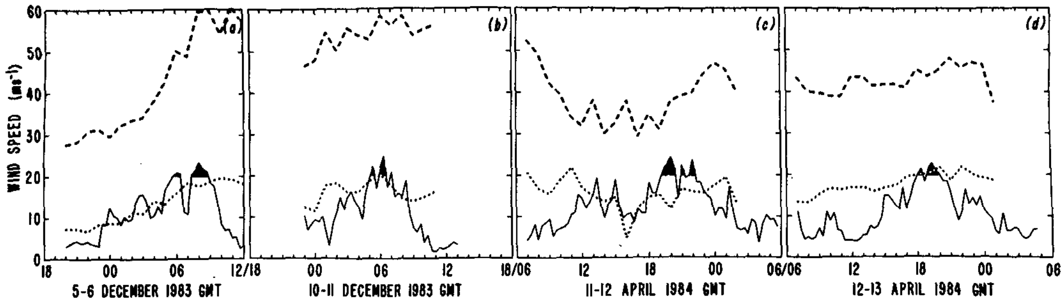

FIG. 3. As in Fig. 2, except (a) case III, (b) case IV, (c) case V, and (d) case VI.

at the jet stream and 650 mb level is compared with the surface winds in Boulder in Fig. 4. (We show only the two strongest windstorms and the nonwindstorm cases.) While an upper tropospheric wind direction of between 280° and 315° may be a necessary condition for strong surface winds, as shown in Fig. 4c, it clearly is not a sufficient condition. Note also that these results confirm the favorable location of the Lay Creek profiler for sampling the airflow upstream of Boulder during windstorm events.

Another parameter which shows some correlation with the occurrence of windstorms is the vertical wind shear. In addition, patterns can be found in the variation in the upper tropospheric wind shear which seem to be at least weakly associated with the onset of strong downslope winds. The normalized direction and magnitude of the lower (650–500 mb) and upper (500 mb–tropopause) level shear in the two strongest and the nonwindstorm cases are plotted in Figs. 5 and 6, respectively. In these figures, the direction of the mean wind in each shear layer has been subtracted from the direction of the shear vector; this prevents the changes in shear direction from being dominated by changes in the mean wind direction (due to the increase in wind with height).

The changes in the lower level shear (Fig. 5) do not appear to be well correlated with the changes in the surface windspeed in Boulder. In contrast, as shown in Fig. 6, there is a distinct tendency in both of the downslope wind cases for the magnitude of the upper-level shear to decrease, and the normalized shear direction to approach zero a few hours before the surface winds reach maximum strength. In the nonwindstorm case, the magnitude of the upper-tropospheric shear is never reduced to the low values which were observed in the first two cases, but the normalized shear direction does pass through zero without triggering high surface winds. The behavior of the upper tropospheric shear in the four weaker downslope wind cases is shown in Fig. 7. The relationship between the shear magnitude and the surface windspeed in cases III and V is roughly similar to that observed in the stronger cases (I and II); however, cases IV and VI are considerably different. The physical significance of this reduction in the upper tropospheric wind shear will be discussed later in connection with our attempt to diagnose windstorms on the basis of the nonlinearity parameter Nh/U.

3. Analysis based on stability and profiler data

In this section we will attempt to refine the diagnosis of upstream conditions leading to the development of downslope winds by using the wind data in combination with stability data calculated from rawinsonde observations at Grand Junction and Lander. We will test the hypothesis of Klemp and Lilly (1975) that strong downslope winds are associated with large amplitude mountain waves produced by partial reflections

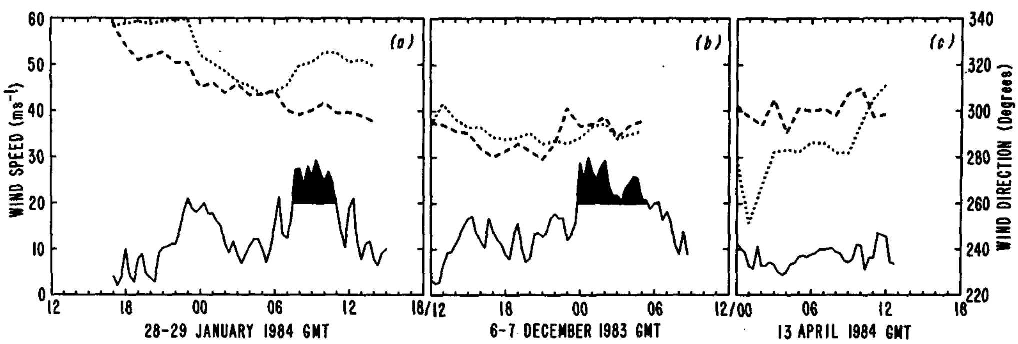

FIG. 4. As in Fig. 2, except the upper air parameters are the 650 mb wind direction (dotted) and the wind direction at the jet stream level (dashed).

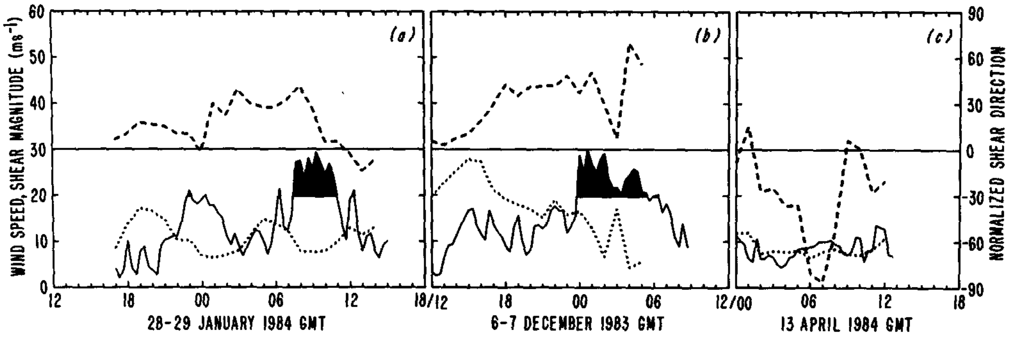

FIG. 5. As in Fig. 2, except the upper-air parameters are the magnitude (dotted) and normalized direction (dashed) of the 650–500 mb wind shear.

from regions where there is an abrupt change in the Scorer parameter. Since the most significant change in the Scorer parameter occurs at the tropopause, they suggest that the most important criterion for strong waves is that there should be a total phase shift of $n\pi$ radians between the ground and the tropopause. A measure of this phase shift may be defined as

$$\phi = \frac{1}{2\pi}\int_{z_0}^{z_T} l dz, \qquad (2)$$

where z_T and z_0 are the heights of the tropopause and the ground (or the top of a layer of stagnant air blocked upstream of the mountain), respectively. The optimal tuning criterion is $\phi = 0.5, 1.0, 1.5$, etc. Following the analysis of Klemp and Lilly (1975), it was assumed that air below 700 mb on the upstream side of the mountain was blocked, so $p(z_0) = 700$ mb.

Equation (2) was evaluated every hour by numerical integration. The temporal resolution of the wind profiler data was one hour, but rawinsonde derived stability data was available only every 12 hours, so it was necessary to interpolate the latter quantity in time. Although the interpolation introduces some error into the computation of ϕ, this error is generally not serious because the variation of the mean tropospheric windspeed in the earth's atmosphere is considerably larger than the variation of the mean tropospheric stability, and ϕ depends only on integrated properties of N and U over the entire depth of the troposphere.

The simplest interpolation scheme would be to construct hourly soundings by linearly interpolating the temperature at each altitude in the new sounding from the temperatures at the same altitude obtained from the 12-hourly rawinsonde data. However, this procedure tends to excessively smooth the sounding structure and to eliminate any stable layers which may be present in the observed data. We chose an alternative procedure, which is based on the fact that regions of strong stability tend to follow isentropic surfaces. Thus, it was generally possible to associate each stable layer in a particular sounding with another stable layer in a previous or subsequent sounding at approximately the same potential temperature θ_{ref}. Then, it was possible to approximate the location of this stable layer at some intermediate time from an estimation of the height of the θ_{ref} isentropic surface. This approach was implemented as follows. Each sounding was approximated by a series of layers having the property that the stability in each layer was constant and equal to the mean stability of the corresponding layer in the observed data. (A good fit for an entire sounding ordinarily required between two and five layers.) Then, each layer was identified, as best possible, with its counterpart in the

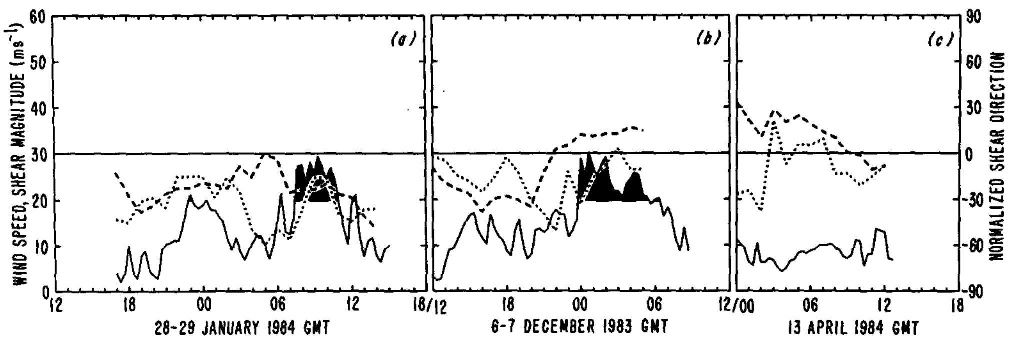

FIG. 6. As in Fig. 2, except the upper air parameters are the magnitude (dotted) and normalized direction (dashed) of the 500 mb–tropopause wind shear.

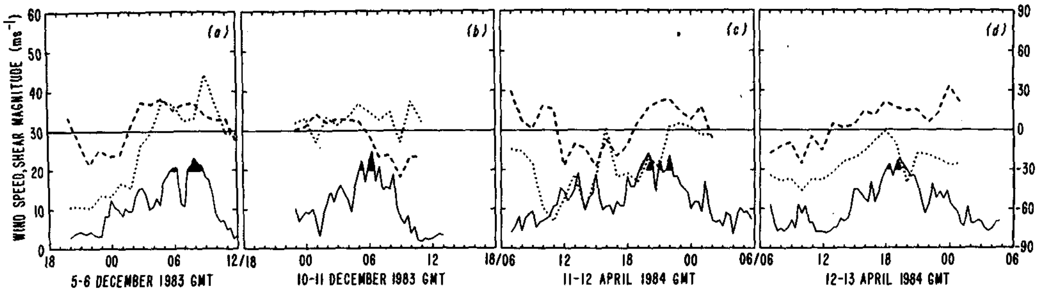

FIG. 7. As in Fig. 6, except for (a) case III, (b) case IV, (c) case V, and (d) case VI.

subsequent sounding, and the potential temperature and height of the tops and bottoms of these layers were linearly interpolated in time to obtain a new sounding.

One additional problem that accompanied our attempt to combine the rawinsonde and wind profiler data sets was that both rawinsonde stations are located a considerable distance from the site of the wind profiler (Fig. 1). In order to avoid excessively smoothing the stability data, we elected not to interpolate the rawinsonde data in space. The phase shifts are calculated independently for both Grand Junction and Lander. Note that the distance between Lander and Craig is approximately twice the distance between Grand Junction and Craig; Grand Junction is also considerably closer to Boulder. Thus, the phase shifts calculated from the Grand Junction stabilities are likely to be more representative than those which were obtained from stabilities at Lander.

When working with observational data, previous investigators have usually neglected the term in the Scorer parameter (1) involving the curvature of the wind profile, so that (2) becomes

$$\phi^* = \frac{1}{2\pi}\int_{z_0}^{z_T}\frac{N}{U}\,dz. \qquad (3)$$

The primary reason for this approximation is that it is difficult to accurately calculate the second derivative of the windspeed from rawinsonde data. (Some authors seem to have erroneously assumed that this term is small.) Since the wind profiler data is reported on a fine mesh (at 290- and 870-m vertical resolution), it would now appear to be feasible to evaluate the full expression for the Scorer parameter. It was surprising to find that inclusion of the curvature term often produces a negative l^2. The data need not be particularly noisy to accomplish this. Consider an example in which $N = 0.01$ s^{-1}, $U(z_{\text{ref}} - 300 \text{ m}) = 30$ m s^{-1}, $U(z_{\text{ref}}) = 29.5$ m s^{-1}, and $U(z_{\text{ref}} + 300 \text{ m}) = 30$ m s^{-1}. Then

$$\frac{N^2}{U^2} = 1.15 \times 10^{-7} \text{ m}^{-2}, \quad \frac{1}{U}\frac{d^2U}{dz^2} = 3.77 \times 10^{-7} \text{ m}^{-2},$$

and the Scorer parameter is dominated by the curvature term. The curvature term is clearly sensitive to small-scale fluctuations in the wind field; however, since the windspeeds used in the evaluation of l are supposed to be representative of a mean state, those small-scale variations which represent perturbations about the mean state should be neglected. An attempt was made to remove the small-scale structure by applying a numerical smoother to the windspeed data, but even after smoothing, the calculation of l was still overly sensitive to the curvature of the wind field (as might be expected from the previous numerical example). This problem was aggravated by the fact that there was a small, but systematic, difference between the profiler windspeeds reported on the 290- and 870-m meshes. Therefore, we will limit the present discussion to calculations based on the approximation (3). The results obtained using the full form of (2) on the smoothed data are similar to those presented here, although the temporal fluctuations in the phase shifts are noisier.

The phase shift calculations for the two strongest windstorm cases and for the nonwindstorm case are shown in Fig. 8. In case I (Fig. 8a), the phase shift drops dramatically to approximately 0.5 four to five hours before the onset of the strongest surface winds. If allowance is made for the time lag required for air parcels to travel to Boulder from Lay Creek or Grand Junction, the transition to an optimally tuned troposphere appears to be well correlated with the onset of the strongest downslope winds. Notice, however, that the earlier windy period around 00Z is not closely associated with any particular period of tuned atmospheric structure. Prior to this earlier period, ϕ^* decreases rapidly, passing through the tuned values of 1.5 and 1.0, and the detuned value of 1.25.

In case II (Fig. 8b), the temporal variation of ϕ^* is considerably smaller; it decreases slowly with time, remaining near the optimal value of 0.5 for a long period before the storm. Note in particular that there is no point at which a rapid change in ϕ^* can be identified with the abrupt onset of strong downslope winds in Boulder. In this case, there is little suggestion that changes in the tuning of the troposphere led to the development of a windstorm. The situation in the nonwindstorm case (Fig. 8c) is similar to that in Fig. 8b in that there is a long period during which the surface winds are light while ϕ^* is approximately 0.5. However, in this case strong winds never develop. (Recall that

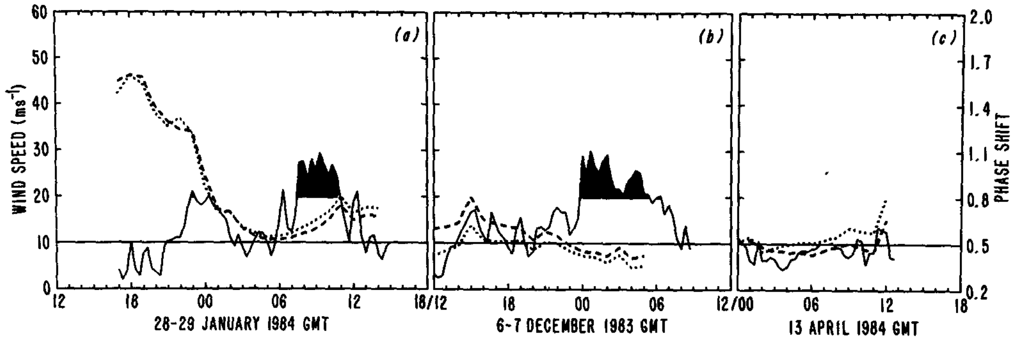

FIG. 8. As in Fig. 2, except the upper air parameters are the vertical phase shifts across the troposphere calculated from the profiler winds, and the stabilities observed at Grand Junction (dashed) and Lander (dotted).

the general characteristics of the air flow appear favorable for mountain waves.)

The results for the four weaker downslope wind events are shown in Fig. 9. Cases III and V (Figs. 9a,c) resemble the behavior in case I, in that there is a substantial decrease in ϕ^* three to four hours prior to the onset of the strong downslope winds. Although there is some difference between the phase shifts calculated using the Grand Junction and Lander stabilities, the average value of ϕ^* (which could be interpreted as the result of spatially interpolating the stability data from Grand Junction and Lander to Lay Creek) is approximately 0.5 during, and immediately prior to, the strongest winds. On the other hand, cases IV and VI (Figs. 9b,d) resemble case II in that ϕ^* shows little variation that can be associated with the development of the winds, and remains near its optimal value for a long time before and after the period of strongest winds.

In summary, the fact that a phase shift of approximately one-half vertical wavelength was associated with every windy period, except the one centered at 00Z in case I, suggests that the optimal tuning criteria may be a necessary condition for the development of strong downslope winds. However, the tuning criteria does not appear to be a sufficient condition, even when the general character of the cross mountain airflow seems favorable for the development of mountain waves. Cases II, IV, VI and the nonwindstorm case provide examples in which $\phi^* = 0.5$ and there is good cross-mountain flow for considerable periods during which the surface winds are weak. This suggests that an additional factor must be important which has not been included in our analysis.

According to the theory of Peltier and Clark (1979), this factor would be wave breaking. Although the parameter Nh/U can be used to determine whether breaking will occur when the windspeed and stability are constant in the upstream flow, there is no equivalent, rigorously correct way to predict the development of breaking waves when there are realistic vertical variations in N and U upstream. In the absence of more explicit theoretical guidance, it seems reasonable to attempt to diagnose breaking from the integrated average of Nh/U:

$$\phi^{**} = \frac{h}{z_T - z_0} \int_{z_0}^{z_T} \frac{N}{U} dz, \tag{4}$$

where z_T is the tropopause height and z_0 the height of the ground, or the top of a layer of stagnant air blocked upstream of the mountain. This is identical to (3), except for the factor $2\pi h/(z_T - z_0)$ which is almost a constant. (Note that z_T varies slightly with time.)

Although (4) appears to be the most straightforward way to attempt to assess whether breaking will occur in those cases where N and U are not constant with height, another alternative would be to evaluate Nh/U at some particular level. This second possibility produces rather noisy results, since large, abrupt changes

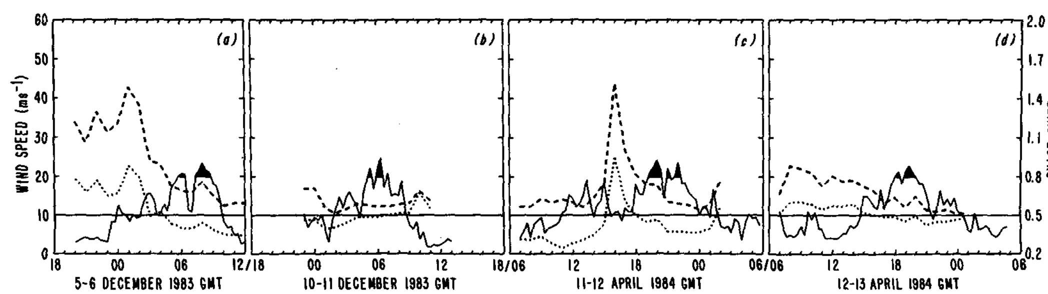

FIG. 9. As in Fig. 8, except for (a) case III, (b) case IV, (c) case V, and (d) case VI.

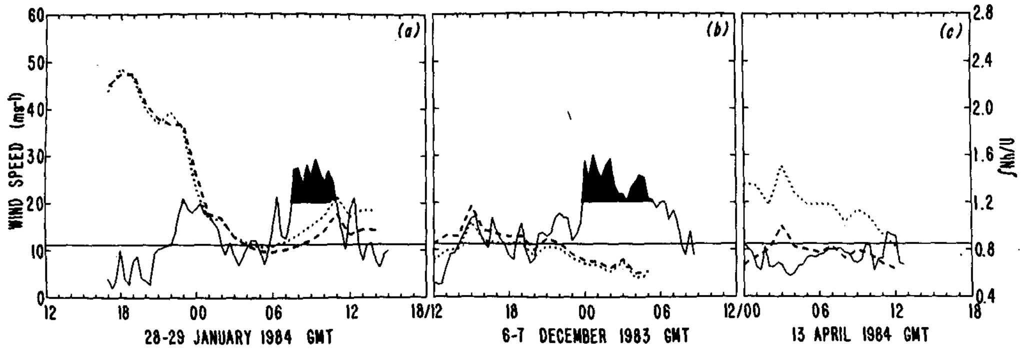

FIG. 10. As in Fig. 2, except the upper air parameters are the integral average of Nh/U across the tropopause, calculated from the profiler winds, and the stabilities observed at Grand Junction (dashed) and Lander (dotted).

in N occur when inversions and stable layers ascend or descend through the calculation level. Because there is little reason to apply a criterion which is derived from the case where N is constant in a manner which is highly sensitive to the height of an inversion layer, we will only consider the integral form (4).

The time dependent behavior of (4) is shown in Fig. 10, for the two strong cases and the nonwindstorm case, and in Fig. 11 for the four weaker cases. If the mountain profile was a symmetric bell shape or a single sine wave, and N and U were constant, wave breaking would occur whenever Nh/U exceeded 0.85; we have, therefore, indicated the 0.85 level with a horizontal line in Figs. 10 and 11. Although 0.85 can be considered a nominal threshold value, it is more important to examine the trend exhibited by the function (4) prior to each wind event. Notice that with the single exception of case V, there is a distinct tendency for the integrated value of Nh/U to decrease prior to the onset of the winds, suggesting that the atmospheric structure becomes less favorable for wave breaking prior to the windstorm. This is in contrast to our earlier results, showing a decrease in the upper tropospheric wind shear prior to the onset of most storms (Figs. 6 and 7), which would tend to support the importance of wave breaking. Given that some mechanism is available to produce a large amplitude wave, the decrease in shear increases the chances of developing a breaking wave in the upper troposphere. It is possible that (4) simply does not provide a suitable formula for the diagnosis of wave breaking in situations where N and U vary substantially with height. Indeed, recent numerical simulations by Durran (1986) suggest that the mechanisms which are responsible for the growth of large amplitude waves in multilayer atmospheres are strongly dependent upon details of the layered structure which would not be reflected in (4).

We have seen that neither of the formulas (3) or (4) provide a particularly accurate way of diagnosing the windstorms in this data set. As discussed previously, the only parameters which affect the airflow in the dry inviscid case are the geometry of the mountain (which is fixed), the windspeed and the stability. The Lay Creek profiler provided an extensive set of wind data collected hourly at a location directly upstream of Boulder. However, stability data were only available at 12-hourly intervals from less ideally located rawinsonde stations. While this stability data may be adequate for estimating the vertical phase shift between the ground and the tropopause (an integral property of the mean tropospheric N), it can not be expected to accurately determine the location and strength of inversions. This may be a serious deficiency since the hydraulic jump theory of downslope winds suggests that the location and strength of upstream inversions should have a critical influence on the character of the flow.

Both Colson (1954) and Brinkmann (1974) have noted that inversions or stable layers are usually present

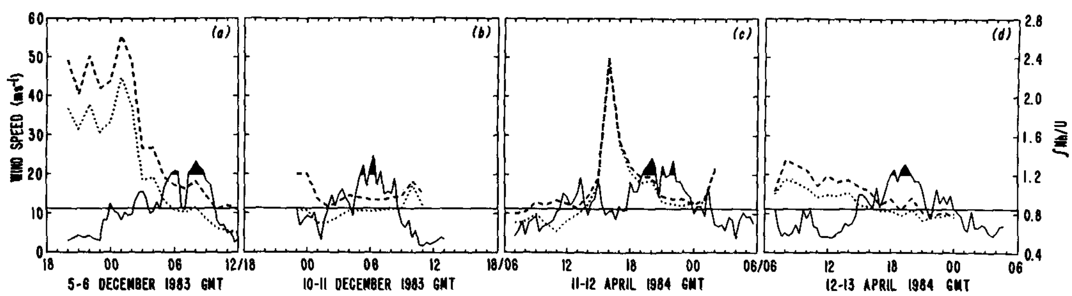

FIG. 11. As in Fig. 10, except for (a) case III, (b) case IV, (c) case V, and (d) case VI.

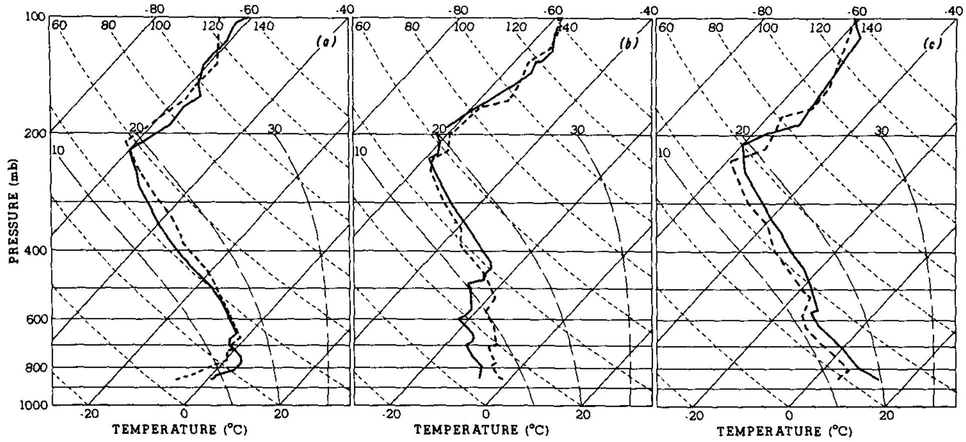

FIG. 12. Grand Junction soundings observed near the time of highest surface winds. The earlier sounding is shown as a solid line; the later one is dashed. (a) Case I: 00 and 12 GMT, 29 January 1984; (b) case II: 12 GMT, 6 December and 00 GMT, 7 December 1983; (c) case VII: 00 and 12 GMT, 13 April 1984.

in the upstream flow near the mountain-top level during strong mountain wave and downslope wind events. The Grand Junction soundings prior to and after the period of highest surface winds in the two strongest cases may be compared with the soundings for the nonwindstorm case in Fig. 12. In both downslope wind cases, there are elevated inversions near mountain-top level at both Grand Junction and Lander, with the stronger stable layers generally observed at Grand Junction. This was also true for the soundings taken before the weaker cases III–VI (except that inversions were only observed at Grand Junction in case V). However, in the nonwindstorm case, there is only a weak inversion at Grand Junction (Fig. 12c) and no trace of an inversion at Lander. Notice also that the lower tropospheric stability is very weak in this last case. One might well hypothesize that the missing element which prevented the development of high winds in case VII was the lack of a strong inversion near mountain top level in the air mass upstream from Boulder.

4. Conclusions

The airflow upstream from Boulder, Colorado has been examined during downslope wind events in an attempt to determine parameters which display a characteristic behavior associated with the onset of the strong surface winds. The parameter set was limited to those which could be calculated from high-resolution wind profiler data and coarse- (temporal) resolution rawinsonde stability data. Both the wind direction and the vertical phase shift across the troposphere appear to have some utility in windstorm prediction in that during each windy period, the wind direction was between 280° and 315°, and the phase shift approximated the optimal value of 0.5. However, there were many instances in which the phase shift and wind direction were optimal, but no downslope winds occurred. A third parameter which appeared to be related to the development of strong surface winds was the shear between 500 mb and the tropopause. In both of the stronger cases, there is a pronounced reduction in the magnitude of the upper-tropospheric shear a few hours prior to the onset of intense surface winds.

However, these parameters do not seem to be sufficient to completely distinguish between windstorm and nonwindstorm events. An additional factor, which is likely to be useful in forecasting downslope winds, is the height and strength of inversions in the upstream flow. The twelve-hourly rawinsonde reports from Grand Junction and Lander suggest that inversions may have been present near the mountain-top level, but it was impossible to estimate their temporal variation from the available data.

An unavoidable shortcoming of this study is the small number of cases examined, and in particular, the lack of a very strong downslope windstorm among the data. The Lay Creek profiler has been taken out of operation, so at present, there is no opportunity to expand the data set. If operation of the Lay Creek profiler is resumed, it would be highly advantageous to add a third vertically pointing beam which could be used to determine the location of stable layers.

It is interesting to consider the extent to which the results of this study support each of the three theories which have been proposed to explain the occurrence of strong downslope winds. The fact that phase shifts close to the optimal value of 0.5 were calculated for periods when the surface winds were strong, and also for nonwindstorm cases with good cross-mountain flow, suggests that while partial reflection from different atmospheric layers may play a role in the production of strong winds, changes in the tuning properties of these layers are not, by themselves, sufficient to trigger windstorms. The fact that a reduction in the upper tropospheric shear was observed shortly before the two strongest cases suggests that wave breaking may play an important role in the development of high winds, especially in the stronger cases. However, we also found that, prior to onset, the integrated value of Nh/U tends to decrease from values which would indicate the presence of overturned waves to values which would not support overturning. This either suggests that breaking is not important, or (more likely) that there exists a mechanism capable of triggering breaking waves when the upstream windspeed and stability vary with height, which is not adequately diagnosed by the integrated value of Nh/U. The fact that no clear triggering mechanism was distinguishable in the available data suggests that some factor involving changes in the vertical structure of the stability might be important. This in turn suggests that hydraulic mechanisms may play an important role in the development of downslope winds, since the character of the airflow predicted by hydraulic theory is a strong function of the height and strength of upstream inversions.

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