Direct Measurement of Large-Scale Vertical Velocities Using Clear-Air Doppler Radars

G. D. NASTROM

Control Data Corp., Minneapolis, MN 55440

W. L. ECKLUND AND K. S. GAGE

Aeronomy Laboratory, NOAA, Boulder, CO 80303

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ABSTRACT

Radars that can make wind measurements in the clear air are expected to play an increasing role in meteorological observing systems in the future, especially for horizontal wind measurements. This paper considers the prospects for using these radars, which are sometimes called wind profilers, to also measure the large-scale vertical velocity. Unfortunately, all radars for which vertical velocity data are available at this time are located in or near mountains, where standing lee-wave effects often make the data representative of only small-scale features. Confining attention to those times when lee wave effects are not expected, case-study comparisons of the existing radar data with indirectly computed synoptic-scale motions suggest that time-averaged radar data are representative of large-scale features smaller than the synoptic scale, perhaps more aptly termed subsynoptic-scale features. Results from a three-station radar network in France show that the time-averaged vertical velocities are usually nearly the same at all stations, although there are some differences, and suggest that the spatial scale of the flow features they represent is greater than 50 km. Over a long-term average, the net influence of lee wave effects at mountain sites is small, and radar measurements appear to be useful for climatological studies of vertical velocity in large-scale circulation systems.

1. Introduction

Doppler radars capable of obtaining winds by measuring the motion of small irregularities in the refractive index are being used at an increasing number of sites over the globe. These VHF or UHF radars can operate continuously in time, providing vertical profiles of one, two, or all three components of the wind vector in the troposphere and lower stratosphere (and mesosphere for large radars). They are thus often called ST (or MST) radars or simply wind profilers. First measurements by this technique were reported by Woodman and Guillén (1974) about a decade ago; since their pioneering work ST radars have been used to measure the horizontal winds at about a dozen locations, and there are reasons to anticipate even wider use of them in the future for both research and operational purposes. A mesonetwork of ST radars is being used as part of the Prototype Regional Observing and Forecasting System (PROFS) research program in Colorado (Strauch *et al.*, 1984; Shapiro *et al.*, 1984) and an even more extensive network of wind profilers is being considered for the upcoming Stormscale Operational and Research Meteorology (STORM) project (UCAR, 1983). For operational purposes, Larsen (1983) has already demonstrated that radar-measured horizontal winds can be used for synoptic analysis. The general outlook for operational wind profiling using ST radars has been discussed by Balsley and Gage (1982), and the potential economic impact of a nationwide dense network of ST radars on airline operations has been pointed out by Carlson and Sundararaman (1982). Clearly, ST radars can be expected to play a significant role in future meteorological observations based on their horizontal wind measurement capabilities alone. Their impact may be even greater if other capabilities, such as vertical wind measurements, can be fully exploited. The purpose of this paper is to consider the question: "Are the average vertical velocities measured overhead by an ST radar related to the large spatial-scale vertical motions associated with the synoptic scale?"

Most previous studies of the radar-measured vertical velocity have dealt with its short-term variability. Larsen and Röttger (1982) reviewed the uses of ST radars for synoptic meteorology, including their ability to monitor continuously the vertical velocity over a station, and they presented the observed vertical motions during one case study of the passage of a warm front over the Sounding System (SOUSY) radar in Germany. Their data clearly show upward motions ahead of the front and downward motions behind the front, and it appears the average velocities in each sector are on the order of 10 cm s^{-1}. This pattern is consistent with classical models such as those given by Palmén and Newton (1969), although Larsen and Röttger hesitate to interpret their velocities as due to the synoptic-scale flow pattern. Indeed, it is difficult to verify the vertical velocity measurements made by clear air Doppler radars for a variety of reasons discussed below.

Of course, the most straightforward way to determine whether ST radars properly measure the vertical air motion would be to compare radar data with an accepted standard measurement technique. Unfortunately, in the case of vertical velocity there is no standard technique routinely available. Aircraft, balloons, and even towers have been used to measure vertical air motions for specific projects, but it is inconvenient or expensive to employ them on a large scale. Thus, as Panofsky (1951) wrote over three decades ago: "The vertical velocity is the only component of the air velocity vector not generally recorded . . . but the average vertical velocity over large areas (10^{14} cm^2 or more) must be estimated indirectly." The situation is still the same, and we are forced to estimate the large-scale vertical motion indirectly. Nastrom (1984) presented the results of a case study comparing radar-measured vertical winds with those computed from the indirect equations and radiosonde data, and found that ST radar measurements reflect the synoptic-scale vertical velocity under certain conditions. In this paper we examine more fully the extent to which time-averaged, single-station radar measurements of vertical velocity are representative of large spatial-scale vertical motions.

2. Observed character of vertical velocity fluctuations and the influence of topography

As a first step, it is worthwhile to examine a sample of the available vertical velocity observations. Figure 1 shows time series of fifteen-minute averages of the vertical velocity at Platteville, Colorado, for a three-week period in March 1981, from Ecklund *et al.* (1982). Two obvious features stand out in these time series. First, the vertical velocity fluctuates rapidly. Second, these time series are characterized by alternating periods of high and low variance of the vertical velocity, i.e., there are active and quiet periods. Furthermore, the active periods tend to occur with about the same intensity at all altitudes. Similar features in the variability of vertical velocity have been noted at Poker Flat, Alaska (Ecklund *et al.*, 1981); Sunset, Colorado (Ecklund *et al.*, 1982); for three closely-spaced radars near Nîmes, in Southern France (Balsley *et al.*, 1983); and appear to be present in hourly-average data from the SOUSY radar in Germany presented by Röttger (1981). Note that all of these sites are located in or near mountains.

For many purposes the Taylor hypothesis can be used to relate temporal and spatial variations in atmospheric motions. If the Taylor hypothesis were applicable to the vertical velocity fluctuations present in the time series of Fig. 1, then time averaging the data should smooth out the most rapid small-scale variations and a mean value representative of a relatively large spatial scale would emerge. However, during the active periods noted above, the observed vertical velocity variance is dominated by large amplitude gravity waves which do not satisfy the Taylor transformation since they can have long periods and very small spatial scales. These large amplitude gravity waves are associated with topography (Ecklund *et al.*, 1982) and are primarily comprised of lee waves.

The presence of standing lee waves with phases tied to local mountain topography is evident in Fig. 1: e.g., on 26 March the velocity of the lower altitudes is strongly downward for several consecutive hours. Other cases can easily be found. Episodes such as this

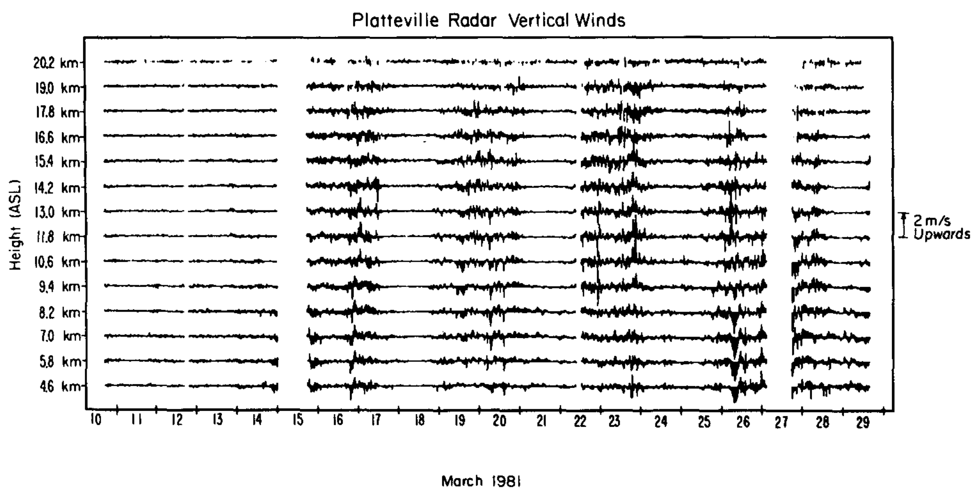

FIG. 1. Fifteen-minute averaged vertical velocities for the period 10–29 March 1981, over Platteville, Colorado. After Ecklund *et al.* (1982).

lead to a very different spectral signature during the active times compared with the quiet times as illustrated by the spectra in Fig. 2 from Balsley *et al.* (1983). The variance during active times is raised relatively much more at periods longer than about 30 minutes than at shorter periods, with the greatest increase at periods near 3 hours. The point is that different processes contribute to the vertical velocity variance during active and quiet periods. The presence of standing waves during the active times severely biases the mean vertical velocity (over time scales of up to many hours) toward smaller spatial scales. Under these circumstances there seems little point in trying to extract a mean value which could be representative of a large spatial area from single-station data. (For longer term averages, however, we anticipate that the phases of the waves will be uncorrelated so that a meaningful measurement of mean vertical motion may still be possible. Long-term averages will be presented later.)

During quiet times we can find no evidence of standing waves by visual inspection of the data (a detailed example of quiet-time data is given by Ecklund *et al.*, 1981). In fact, the spectrum for quiet days in Fig. 2 closely resembles that expected for a spectrum of internal gravity waves. These waves do not have their phase tied with the topography, but rather are traveling waves and their net contribution to the time-averaged vertical velocity over the station is zero. During quiet times the average radar-measured vertical velocity should thus be biased only by the large-scale vertical air velocity, and should give the desired measurement.

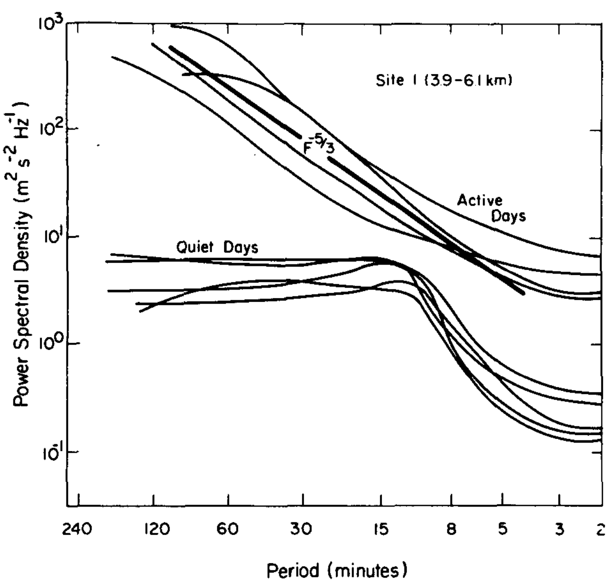

FIG. 2. Representative spectra of the vertical wind fluctuations for active and quiet periods obtained during ALPEX at a site near the mouth of the Rhône River in France (after Balsley *et al.*, 1983).

3. Statistical considerations for time averaging vertical velocities

Since time averaging is clearly necessary to average out the small-scale variability of vertical velocities associated with internal gravity waves, we consider next how much averaging is required from a statistical point of view to obtain the desired measure of large-scale vertical motion.

The statistical confidence of time-averaged data can be estimated by the standard error of the mean (SE), given by $SE = C\sigma\sqrt{N}$, where σ is the standard deviation of the data and N is the number of independent data points used. Due to autocorrelation there is a typical minimum time interval between independent observations, regardless of the sampling frequency (Leith, 1973). Nastrom and Gage (1983) estimated that the time between independent observations of vertical velocity during quiet periods is about 9 minutes. The standard deviation during quiet periods is typically near 10 cm s^{-1}, independent of altitude, and changes very little with the averaging period used when the averaging period is over a few hours in length. The C term is a statistical factor which depends on the confidence limit imposed and the number of observations used; at the 90 percent limit for a large number of observations C is about 1.65. With C and σ fixed, the expression for SE can be used to determine the value of SE for specified N, or to determine N for specified SE. We have arbitrarily chosen to set $SE = \pm 2$ cm s^{-1}, thereby requiring 9-hour averages of the radar data. Note that SE decreases only slowly if longer averaging times are employed (e.g., $SE = \pm 1.83$ cm s^{-1} for 12-hour averages, ± 1.5 for 18-hour averages, and ± 1.3 for 24-hour averages), and the statistical confidence of the results given below is not very sensitive to the length of the averaging interval. This estimate applies only for quiet periods, as σ is larger during active periods. Also, if large-amplitude topographically-forced gravity waves are present, the number of independent samples will be reduced substantially due to their long autocorrelation time so that SE will be larger for the same averaging period. Thus, even "weak" lee wave activity could compromise the measurement of large-scale vertical motion.

The spatial scale corresponding to these 9-hour averages is unknown, and one of our objectives is to see if it approximates the synoptic scale. We expect that it will turn out to be representative of a scale smaller than the synoptic because the rawinsonde data upon which the NMC analyses are based are spaced about 400 km apart on the average. Thus, the "synoptic-scale analyses" will only begin to show features with scales near about 800 km. Assuming that the Taylor hypothesis is valid and taking a mean horizontal flow speed of 15 m s^{-1}, 486 km of air stream over the radar in 9 hours, and unless the spatial correlation of vertical velocity falls off very slowly with distance, we should not expect our results to be representative of scales greater than 400–500 km, at most. The observed relationship of the synoptic- and subsynoptic-scale vertical motion fields is discussed below.

4. Comparisons of vertical velocities

a. Data

The ST radar measurements at Platteville, Colorado, and from a site south of Nîmes, France, have been discussed in detail by Ecklund *et al.* (1982), and Balsley *et al.* (1983), respectively. Nine-hour averages of the radar data are used. The original data have a time resolution of from 1 to 5 minutes.

Indirect estimates of the large-scale vertical velocity were made using the adiabatic, kinematic and quasi-geostrophic omega equation methods. The equations used are given in the Appendix. These are not the only methods available for estimating vertical velocity, but were chosen because they are commonly used. Intercomparisons and discussion of different vertical velocity methods dates back at least to World War II (e.g., Fleagle *et al.,* 1945; Panofsky, 1946), and is still an active topic of research. Indeed, the large number of papers on this topic over the years, with sometimes varying results, helps to illustrate the importance and difficulty of estimating the vertical velocity. Recent results by Wilson (1976) and Smith and Lin (1978) tend to favor the kinematic method, although the omega equation was found useful by Heflick and Fors (1979). Perhaps the most important point to mention here is that each of the indirect methods has inherent uncertainty and errors. Some of these errors are due to assumptions made in deriving the method (such as the assumption of adiabatic flow), while others are due to inaccuracies or inadequacies of the input data. Most importantly, the magnitude of the derived vertical velocity is quite sensitive to the spacing of the input data. Belt and Fuelberg (1982) used AVE-SESAME I data to compare vertical motions derived at 500 mb using subsynoptic- and synoptic-scale data sets. For the same meteorological conditions they found vertical motions derived using the subsynoptic-scale data were as much as 6 cm s^{-1} larger than those derived using the synoptic-scale data. For example, in one case upward motion exceeding 6 cm s^{-1} in a closed center over Arkansas in the subsynoptic-scale analysis is barely perceptible in the synoptic-scale analysis. Evidently, subsynoptic-scale vertical velocities are greatly smoothed in the analysis of derived vertical motion when only synoptic-scale observations are available. Subsynoptic-scale features are a part of the atmospheric flow, and while time averaging the radar data should help smooth them out we should not be surprised to find that the amplitudes of vertical velocity profiles are exaggerated (upward or downward) in single-station radar data compared with synoptic-scale estimates.

b. Case studies

The comparisons of radar data and indirectly computed vertical motions will be confined to quiet periods in the radar data. In general, the quiet periods tend to be associated with rather uninteresting synoptic patterns with small (<2 cm s^{-1}) indirectly computed vertical velocities. We have examined all the quiet-time data for the periods March 1981–October 1982 at Platteville and April–May 1982 for the radars near Nîmes, France. In about 85 percent of the quiet-time cases the radar-measured and indirectly computed values agreed within ±2 cm s^{-1}, the standard error of the mean for 9-hour averages of radar data, nearly as expected. This result is certainly inconclusive, because a prediction of $w = 0 \pm 2$ cm s^{-1} would have produced the same statistical results. In order to compare the radar and indirect vertical velocities in a meaningful situation, we have chosen to present three case studies associated with synoptically interesting flow patterns when the large-scale vertical motion exceeded 2 cm s^{-1}.

As a first case study we consider the series of profiles on 21–22 March 1981 at Platteville (Fig. 3). The synoptic situation at this time showed northeast-

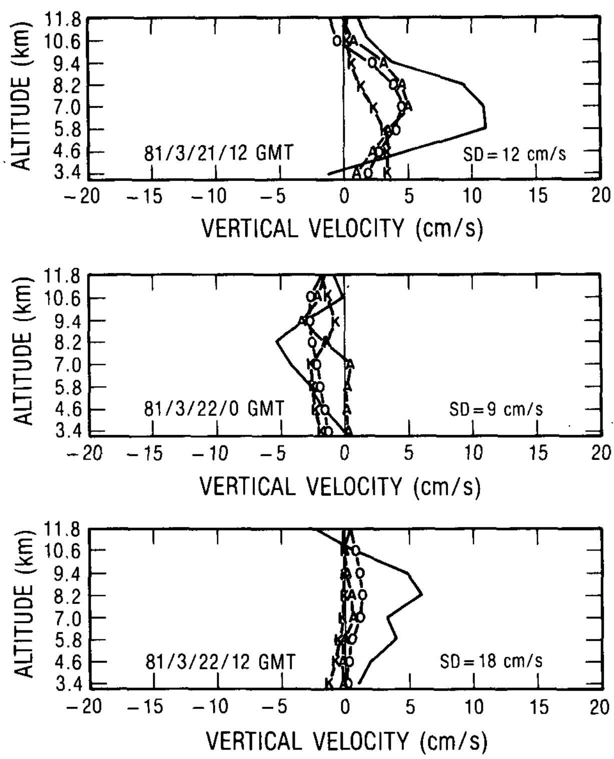

FIG. 3. Profiles of the vertical velocity over Platteville from the ST radar (solid line), and the adiabatic (A), kinematic (K), and omega equation (O) methods.

erly flow aloft over Colorado, with a series of short wave troughs and ridges associated with a closed low over Oklahoma (Fig. 4). In the upper and lower panels of Fig. 3 short wave troughs were approaching Platteville and the vertical velocities were upward throughout the midtroposphere. The radar-measured velocities are generally larger than the indirectly computed values; perhaps because the vertical velocity field contains subsynoptic-scale features, such as the slope of the terrain, reflected in the radar data but not in the NMC-based values. The three-hourly mean vertical velocities over the radar (Fig. 5) near the 500 mb level suggest that the vertical velocity patterns were either of small spatial scale or rapidly propagating or both.

Returning to Fig. 3, the vertical velocities in the center panel are generally downward, although below about 8 km the results from the adiabatic method are near zero or slightly positive. Careful reanalysis by hand of the original radiosonde data failed to provide a significant change in the results for the adiabatic method. The discrepancy in Fig. 3 could arise if the adiabatic assumption were seriously violated. Or, more likely, the time derivative of temperature used in our formulation of the adiabatic method is incorrectly estimated by the 24-hour time differences in this case of a small-scale, rapidly-moving system.

The second case study, also at Platteville, is for 12–14 May 1982 (Fig. 6). During this period a vigorous low moved rapidly out of Arizona into southeastern Colorado, and then began filling and stagnating (Fig. 7). The vertical velocities are upward in the upper three panels of Fig. 6, and then diminish toward zero in the lower two panels. The interesting relatively very large radar velocities near 5 km in the center panel can be traced to an upslope flow condition in the easterly flow aloft over Platteville associated with the low center in southern Colorado. Upslope flow events are subsynoptic scale and are associated with orographically-induced upward motions and heavy precipitation along the eastern slope of the Colorado Rockies (heavy rain and snow fell during this event; Gage and Nastrom, 1985). These results, as in the first case study, suggest that the radar data are representative of a large geographical area, which

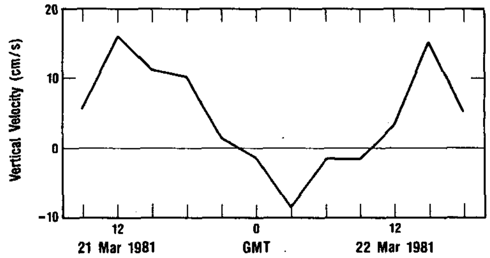

FIG. 5. Time series of three-hourly mean vertical velocities at 5.8 km at Platteville on 21–22 March 1981.

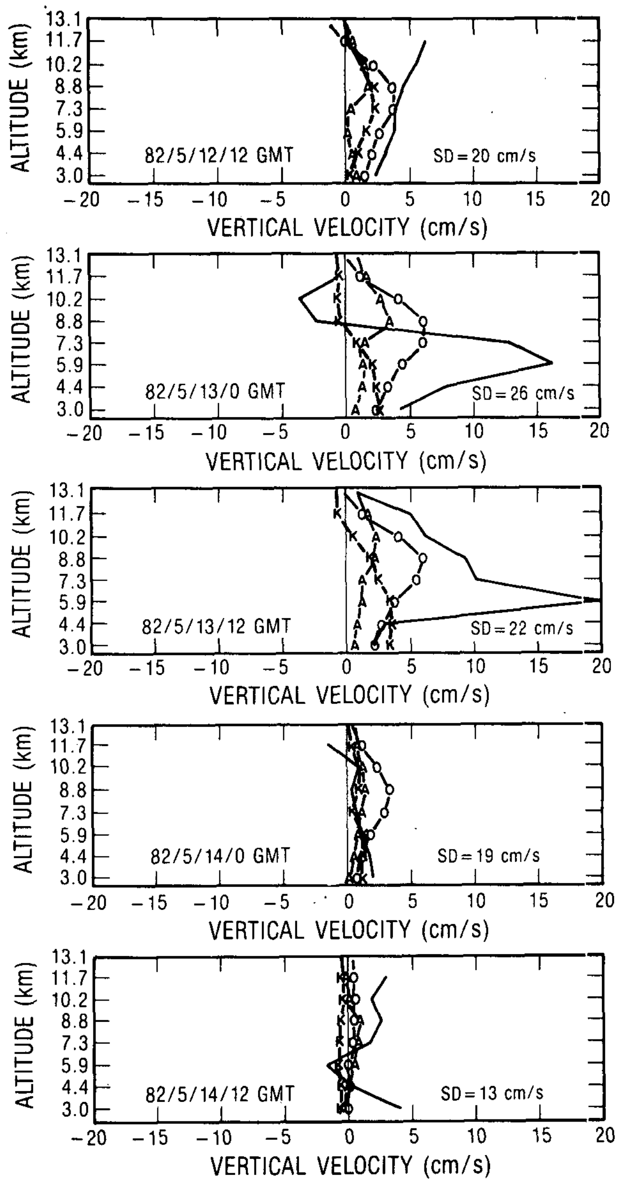

FIG. 6. As in Fig. 3 except for 12–14 May 1982.

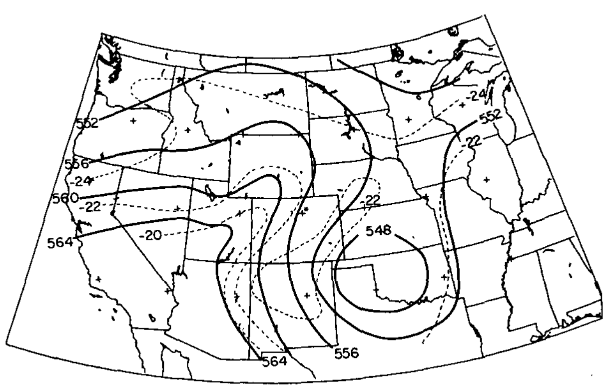

FIG. 4. 500 mb synoptic map for 0000 GMT 22 March 1981.

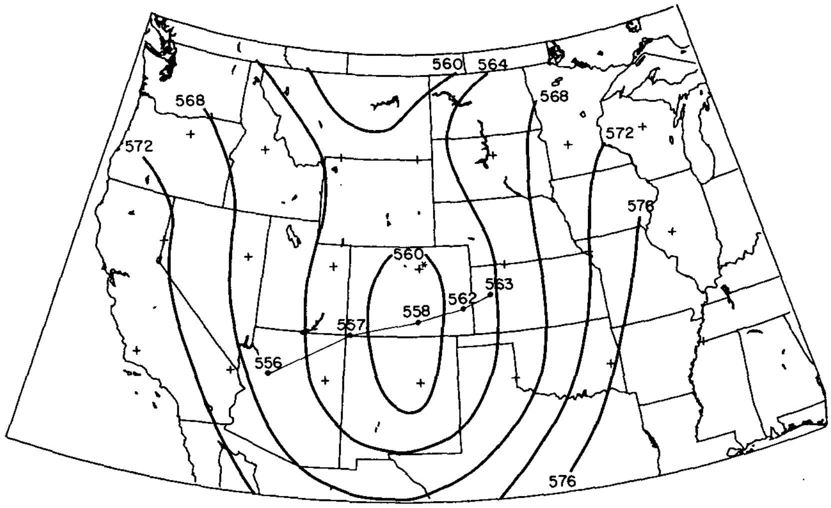

FIG. 7. 500 mb synoptic map for 1200 GMT 13 May 1982. Dots show location and center height of the low pressure center at twelve-hourly intervals.

is perhaps closer to the subsynoptic than to the synoptic scale. In order to judge more accurately which spatial scale the 9-hour averaged vertical velocity data represent it is necessary to employ more than one radar in a network. Ideally, such a network should be located well away from mountains. Unfortunately, the only network of radars which has been used to measure vertical velocities was located in the vicinity of the Alps in southern France.

As part of the Alpine Experiment (ALPEX), three radars were located on the alluvial plain near the mouth of the Rhône River forming the corners of an approximately equilateral triangle with 5 km legs (Balsley *et al.,* 1983). The local terrain was very flat, with the sea south of the radars and the Alps Mountains or foothills in other directions. The radars were operated during April–May 1983, and time series of the vertical velocity data given by Balsley *et al.* (1983) closely resemble the time series in Fig. 1; i.e., there are alternating quiet and active periods. The quiet periods correspond with times of light horizontal winds or flow from the south, off the sea, and the strongest active periods are at times of the mistral when the wind blows strongly from the north.

Average vertical winds for active and quiet periods look entirely different. During active periods vertical wind amplitudes are very large and there is little or no correlation between the three stations. During quiet periods, however, the three radars tend to show agreement. Nine-hour averages of the vertical velocity from these radars during a quiet period in May are shown in Fig. 8, along with the indirectly computed vertical velocities. The synoptic situation was dominated by a blocking ridge near the west coast of Europe (Fig. 9), leading to relatively light winds aloft and southerly flow at the surface. The vertical velocities were generally downward throughout this period. Data for all levels below 12.1 km are included in Fig. 8, although the radar values above about 8 km are of lower statistical quality than those at levels below 8 km. These radars were of relatively low power, and a significant number of the echoes in the 8–12 km range had an unacceptable signal-to-noise ratio and so were not used in forming the means in Fig. 8. The statistical errors above 8 km may thus be up to twice as large as below 8 km. Below 8 km the radar values are usually within 2 cm s^{-1} of each other, and often within 1 cm s^{-1}. However, there are occasionally significant differences among the vertical velocities measured at the three sites, such as on 0000 GMT 14 May. The explanation for these differences is not obvious, and their presence serves to point out how very little is known about vertical velocity and its variability. *In most cases* in Fig. 9, the direction of the radar and the computed vertical velocities agree, although the magnitudes of the radar-measured vertical velocities are often larger than the indirectly computed values; again perhaps due to a subsynoptic-scale effect such as an upslope flow condition on the southern slope of the Alps.

Velocity differences between the three radar sites appear to be random, at least in the brief data sample used here. For example, the mean values over the period 2000 GMT 10 May–0500 GMT 15 May at 5.4 km are -1.7, -1.8, and -1.5 cm s^{-1} for sites 1, 2, and 3, respectively. As these means are not different at the 90 percent confidence level, we tentatively ascribe the day-to-day differences to sampling fluctuations. The average correlation of the time series of 9-hour mean vertical velocities between the pairs of stations is about 0.8. If the shape of the spatial correlation function of vertical velocity were known,

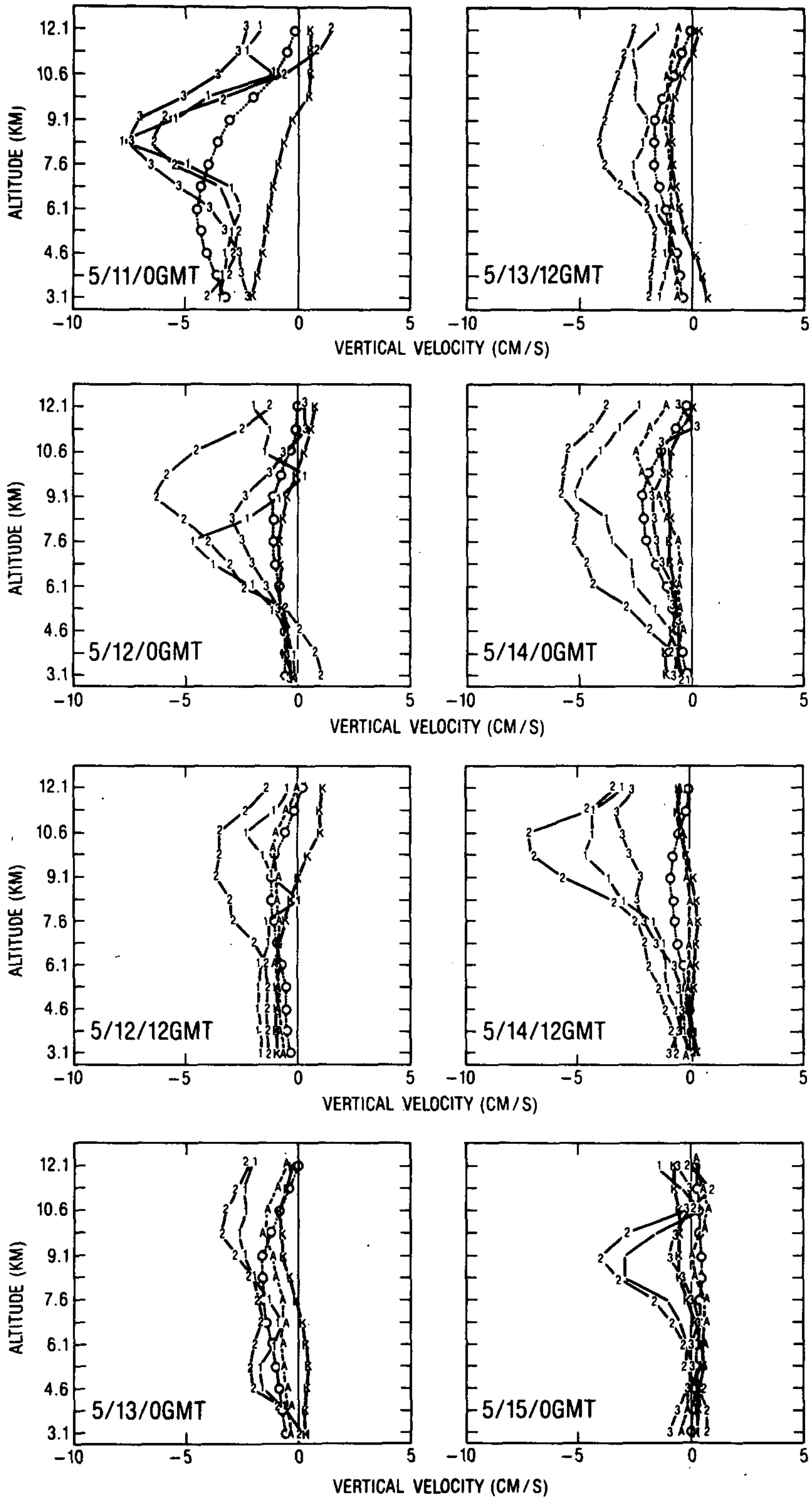

FIG. 8. As in Fig. 3, except radar data from ALPEX, France, for 11–15 May 1982.

we might use this information and follow Leith's (1973) approach to estimate the distance between independent vertical velocity observations. For example, assuming that the spatial correlation function follows an exponential decay with increasing distance, we estimate the distance between independent obser-

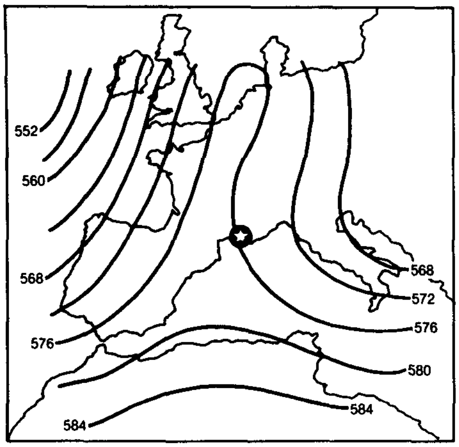

FIG. 9. Synoptic map at 500-mb level for 0000 GMT 14 May 1982. Star is location of ALPEX ST radars.

vations is about 50 km. If other models of the spatial correlation function are assumed, different estimates are obtained, most of them longer than 50 km, suggesting that 50 km is a reasonable lower bound for the distance between independent observations. This value of 50 km can be contrasted with the earlier estimate of 400–500 km for the horizontal scale of the 9-hour averaged vertical motions as discussed in Section 3. Thus there is too much small-scale variability present in the averaged vertical velocities from the three-station network to be consistent with the earlier estimate. A likely source of this small-scale variability of vertical velocity is weak topographically-induced gravity waves. The influence of lee waves is so great that even weak wave activity could produce the complex pattern of small-scale bias that is observed. Another possible source of such variability would be small pointing errors of the individual antennas. Indeed, small random pointing errors would lead to a complex pattern of bias dependent on horizontal wind speed and direction. Clearly, in future experiments great care must be taken to minimize such errors.

c. Long-term averages

The phase and amplitude of standing waves over a mountain site during active periods is a variable function, strongly dependent upon the speed and direction of the winds at the surface and aloft, the distribution of local topographic features, and static stability. Thus, on a long-term (climatological) basis the net contribution of standing waves to the mean vertical velocity may be relatively small, leaving only the contribution due to the large-scale flow. In that case, vertical velocity measurements even at radar sites near or in mountains might provide information on large-scale features for climatological studies. To investigate this possibility we have stratified all available data from Platteville according to the speed of the meridional wind at 300 mb. The meridional wind speed is expected to be correlated with the vertical velocity in a climatological sense because the mean isentropic surfaces in the troposphere slope upward toward the north. Thus, northward moving air should have upward motion on the average, and vice versa. The average radar data shown by the solid lines in Fig. 10 correspond with this expected pattern. The number of 9-hour observation periods averaged together in each of the six categories ($< -20, \ldots, >20$ m s^{-1}) in Fig. 10 is 42, 50, 96, 105, 43 and 33, respectively. Of course, the observed relation between vertical velocity and meridional wind speed could arise by chance if the radar beam happened to be pointed slightly off-vertical and to the south so that the vertical wind measurements were contaminated by horizontal wind. We have considered this possibility and conclude that at Platteville beam pointing errors are not significant.

Average results for the indirect estimates of vertical velocity have been added in Fig. 10 for comparison (results for the kinematic method are nearly the same as for the omega equation at both levels, so are omitted for clarity). The curves for the indirect meth-

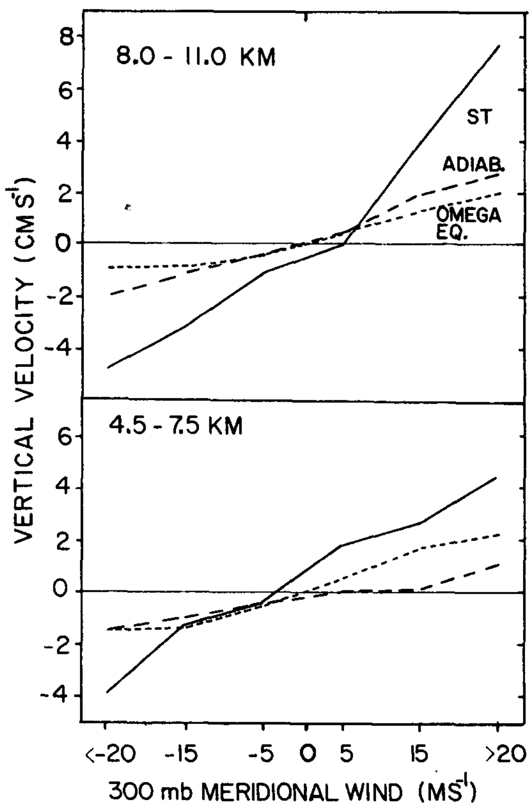

FIG. 10. Mean vertical velocity over Platteville as a function of 300 mb meridional wind speed.

ods also show the expected relationship with meridional wind speed, but in general their amplitudes are less than those from the radar data, primarily for large wind speeds. This discrepancy is probably due to the spatial *smoothening* inherent in the synoptic-scale NMC analyses which causes the indirectly computed vertical velocities to underestimate atmospheric subsynoptic vertical motions, especially in extreme cases. The important point to be drawn from Fig. 10 is that ST radar data appear to be useful for long-term climatological studies of vertical velocity even at a station near the mountains. The meridional wind speed was chosen as the independent variable here because of its well-known relationship with vertical velocity.

5. Summary and conclusions

Recognizing that existing clear-air Doppler radars are located too close to mountains to be ideally suited for measuring large-scale atmospheric vertical motions, we have explored the potential of several existing radars for measuring large-scale vertical motions under "quiet" conditions when the influence of topography should be minimal. The principal analytical technique employed is to compare 9-hour average radar-measured vertical winds with indirect vertical velocities deduced from NMC analyses using the adiabatic, omega and kinematic methods. The 9-hour averaging interval was selected to determine vertical motion ± 2 cm s^{-1} with 90% confidence.

Two case studies using the Platteville radar located in eastern Colorado illustrate the level of agreement obtainable between the average vertical motion observed by the radar and deduced from the indirect methods under these conditions. The two independent measures of vertical motion generally agree with respect to the direction of motion. However, the magnitude of the vertical motions measured by the radar are often significantly larger (by about a factor of 2) than the indirect vertical velocities. This enhancement of the magnitude of directly measured vertical motions is suggestive of subsynoptic influences which in eastern Colorado would include the effect of terrain slope.

Comparison of average vertical velocities measured independently by a network of three ST radars operated in southern France during ALPEX show reasonable agreement during quiet conditions when terrain-induced wave effects are minimal. However, the discrepancies are large enough to suggest that either 1) there is a residual small influence of lee wave effects even under apparently quiet conditions or that 2) small antenna beam pointing errors are leading to a complex pattern of small-scale variability. Nevertheless, day-to-day differences are randomly distributed so that long-term averages appear to lead to meaningful estimates of large-scale vertical motion in a climatological sense.

In addition to the above case-study results, we have made statistical comparisons of directly and indirectly measured vertical velocities with meridional winds. These show, as expected, that upward vertical motion is associated with northward flow. The magnitude of the effect is more pronounced for directly measured vertical velocities (as compared to indirectly determined vertical velocities), once again suggestive of subsynoptic-scale influence.

The results presented here clearly show that ST radars are useful for measuring the large-scale vertical velocity. The spatial scale which these observations represent appears smaller than the synoptic scale, and perhaps should be called the subsynoptic (meso α) scale. Considering the enormous potential benefit of direct measurements of the vertical velocity for operational forecasting and for research purposes, these results are very encouraging. All of the cases considered here were limited to special synoptic situations due to the presence of active periods of large vertical velocity variance induced by mountains. Whether such active periods occur at sites over the plains, far from the mountains, is an open question even though all available evidence links the active periods with the mountains rather than with an inherent meteorological process. Of course, other processes such as convection will undoubtedly at times lead to large variance of the vertical velocity over the plains, but their influence on the large-scale vertical velocity measurement cannot be determined from the data now available. An important next step is to operate a clear-air Doppler radar in a flat-land location to determine whether measurements of the large-scale vertical velocity can, in fact, be made there on a routine basis.

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APPENDIX

Indirect Computation of Vertical Wind

The adiabatic method for computing vertical velocities is based on the first law of thermodynamics, assuming that the flow is isentropic. Vertical velocity in pressure coordinates ω is given by

$$\omega = -(\partial\theta/\partial t + \mathbf{v}\cdot\nabla_p\theta)/(\partial\theta/\partial p), \qquad \text{(A1)}$$

where θ is potential temperature, $\mathbf{v}$ is the horizontal wind, and p is pressure. The time derivative was computed using centered time differences; as data are available only each 12 hours the assumption of isentropic flow may occasionally be seriously violated.

The quasi-geostrophic omega equation is derived from the vorticity equation and the first law of thermodynamics assuming the flow is isentropic and quasi-geostrophic. Other assumptions and a derivation are discussed by Holton (1972). The version of the equation used here is

$$\left(\sigma\nabla^2 + f_0{}^2\frac{\partial^2}{\partial p^2}\right)\omega = f_0\frac{\partial}{\partial p}[\mathbf{v}\cdot\nabla(f_0{}^{-1}\nabla^2\Phi + f) - \nabla^2[\mathbf{v}\cdot\nabla(\partial\Phi/\partial p)], \qquad \text{(A2)}$$

where σ is the static stability parameter constant at each pressure level p, f the Coriolis parameter, and Φ is the geopotential height. Following LeDrew (1980) this equation was solved numerically on an 8×8 grid centered near the radar site with side and top boundary values taken to be zero. LeDrew reports that this grid is large enough that the values at the center of the grid are not very sensitive to the side boundary values assumed. The upper value was set to zero in the absence of other information. The lower boundary values were taken to be the sum of orographic and frictional effects:

$$\omega_0 = \mathbf{v}_e\cdot\nabla p_e - p_e F\zeta_e/2f_0, \qquad \text{(A3)}$$

where subscript e refers to the assumed height of the Eckman layer, taken to be 850 mb over the western United States, and F is the frictional coefficient taken to be 8×10^{-6} s^{-1}.

The kinematic method is derived from upward integration of the equation of continuity

$$\omega = \omega_0 + \sum_i (\nabla\cdot\mathbf{v}_i)\Delta p_i, \qquad \text{(A4)}$$

where ω_0 was computed using Eq. (A3). Vertical velocities computed using Eq. (A4) were adjusted to equal the adiabatic method's value at the 100 mb level via the scheme suggested by O'Brien (1970). The resulting values of ω from each of the above methods were converted to vertical velocity in height coordinates (w) through

$$w = -(RT/pg)\omega, \qquad \text{(A5)}$$

where R is the gas constant, T temperature, and g = 980 cm s^{-2}. Results were then linearly interpolated between pressure surfaces to the heights of the radar data levels.

Estimates of the synoptic-scale vertical velocity were computed with the above methods using NMC gridded analyses of height, temperature and wind. The NMC fields were obtained from NCAR on magnetic tape at 0000 and 1200 GMT daily for the 1000, 850, 700, 500, 400, 300, 250, 200, 150 and 100 mb levels. The methods used to produce these analyses have been discussed by Kistler and Parrish (1982). Also, radiosonde data from Denver and surrounding stations were obtained from the National Climatic Center, Asheville, North Carolina. These were used to make hand analyses in a few cases for more detailed comparison than possible with the highly-smoothed NMC analyses.

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