N91-26622

Chapter 9

Applications of MST Radars: Meteorological Applications

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1. Introduction

MST radars provide information about a broad range of scales. For example, wind measurements provide information about circulations ranging in size from the microscale to the synoptic or planetary scale; spectral width measurements contain information about microscale turbulence within the beam; and reflectivity measurements show the presence of small scale turbulence, as well as larger scale temperature gradients associated with the tropopause or frontal boundaries. In spite of the broad range of scales that can be investigated and the wide range of potential applications of MST radars, I will limit the discussion to those applications that are directly relevant to mesoscale meteorology since the radar technology seems to have found a natural niche in that area. The applications to mesoscale meteorology include using the radars either as research tools to improve our understanding of certain dynamical systems or as part of a network used to provide input data for weather forecasting. The goal of operational meteorology is twofold. The first is to forecast the weather over a variety of time scales, and the second is to monitor conditions so that we can assess the state of the atmosphere at a given time. Weather forecasting is probably one of the most commonly recognized tasks of operational meteorology since virtually all members of the public become users of such a service at one time or another. However, forecasting often includes more than a prediction of the chance of showers and the high and low temperature for the next day. User specific forecasts can include the air stagnation in a metropolitan area, winds aloft for routing aircraft, the strength of onshore winds and the associated beach errosion, or the strength and location of small-scale eddies and the effect of such eddies on chemical spraying in agriculture. The list goes on and on.

The workhorse of the operational observing network has been and still is the radiosonde balloon which provides measurements of pressure, temperature, humidity, and winds up to heights of 16-20 km, i. e., into the lower stratosphere. The average separation between radiosonde stations in the United States and Europe is ~150-500 km depending on the area. In addition to the radiosonde network, a much denser network of surface reporting stations cover the industrialized countries. The latter report the same quantities as the radiosonde provides, but only at the surface, and they also provide information about cloud cover, local precipitation, and tendencies, i. e., temporal changes, of the various quantities.

The station separation and measurement techniques used in the observing network lead to a natural emphasis on synoptic meteorology which is characterized by spatial scales of ~500-5000 km and time scales of 1/2 to 5 days. Increasingly the interest of the national weather services is focusing on problems in mesoscale forecasting (see, e.g., Ray, 1986), first because we do not know very much about most mesoscale dynamical processes and second because most of the severe weather is associated with mesoscale phenomena. The mesoscale is characterized by spatial scales of ~1-500 km and time scales of 1/2 to 12 hr. Satellites have provided important mesoscale data, particularly satellite photographs, but the thermodynamic and wind measurements that the satellites yield generally have poor vertical resolution which is unacceptable.

The seeds of MST radar technology have been sown on a fertile field. The radar wind profiler appears to solve many of the problems related to obtaining mesoscale observations needed for forecasting. The radars easily produce the time resolution, height resolution, and height coverage that are needed. A single radar still does not provide any extended spatial coverage, but the cost of the instruments is relatively low so that the possibility of deploying many of them still exists. The radars only provide direct measurements of the winds, which could be a drawback, but two factors operate in the radars' favor in this regard. The first is a result of geostrophic adjustment theory which predicts that at large scales, the wind field tends to adjust to the imposed pressure field, but at small scales the reverse happens. Therefore, if only one mesoscale quantity can be measured, the wind field is probably the best choice. The second factor again derives from the dynamic relationships and exploits the fact that the wind field is interrelated with the pressure and temperature fields through the balance equation and the vorticity equation. Thus, if the wind field is known over an extended horizontal region, information about the temperature and pressure fields can be derived. The accuracy and usefulness of these techniques still need to be carefully assessed but show great promise for the application of the radar technique to operational mesoscale forecasting.

My intent here is not to provide a complete review of the applications of MST radars to meteorological operations or research or to provide a complete literature review. Therefore, some important work will probably be slighted in my treatment. Rather, I hope to make the reader aware of some of the important issues presently being considered, and some of the potential of the technique for future applications. With this background, I will proceed to describe more specific aspects of the application of MST radar technology to meteorology. The next section deals with the characteristics of the horizontal wind measurements, Section 3 describes how the reflectivity measurements obtained from the radars can be used, Section 4 deals with the vertical velocity measurements from the radars, Section 5 describes quantities that are derived from the radar measurements, Section 6 deals with special characteristics of an operational system, and Section 7 outlines some of the important open questions that remain about the usefulness of the technique.

2. Horizontal Velocity Measurement Capabilities

Interest in using MST radars for studies of mesoscale dynamics derives primarily from the possibility of measuring the horizontal winds over a large height range, with good height resolution and excellent time resolution. The time resolution, in particular, is the main advantage in using radars instead of balloons for horizontal wind measurements.

(a) Precision, Accuracy, and Errors

The wind profiler's capabilities for measuring the horizontal winds are the ones that have been explored most extensively to date. The first experiments that were carried out involved comparisons of the horizontal winds obtained from the radar measurements with the wind data from nearby radiosonde stations. An example of the type of agreement that can be expected is shown in Figure 1 which represents measurements made with the SOUSY-VHF-Radar located in the Harz Mountains in West Germany with wind data from the nearby Hannover radiosonde station (Röttger, 1983). The comparisons have shown good agreement from the outset and have led to a general acceptance of the radars' wind measurement capabilities. Two early studies attempted to quantify the comparisons by using longer time series and by making a better assessment of the magnitude of the errors that could be expected. The first was carried out by Fukao et al. (1982) and was based on 26 days of radar data obtained with the 430-MHz Arecibo radar and the corresponding radiosonde data from the San Juan station. Figure 2 shows the relative locations of the two sites and the trajectories of the balloons launched during the period. The figure also shows that the balloon measurements actually cover a large area during the ascent which can be a problem if small-scale flow features are of interest. The agreement was generally good, although errors varied between ~5 m/s in the troposphere and ~3 m/s in the lower stratosphere. The steady character of the flow in the subtropical region led to a nearly constant pattern in which the ascending balloons first traveled slightly to the east of San Juan, away from Arecibo, and then turned toward the west at higher altitudes. The authors argued that the differences in the lower stratosphere could be explained by errors in the radiosonde observations that occurred when the balloons were far from the launch site and therefore at low elevation angles. At least part of the difference between the two data sets in the troposphere, however, had to be due to variability in the flow.

Jasperson (1982a,b) has carried out numerous twin balloon experiments. Combinations of

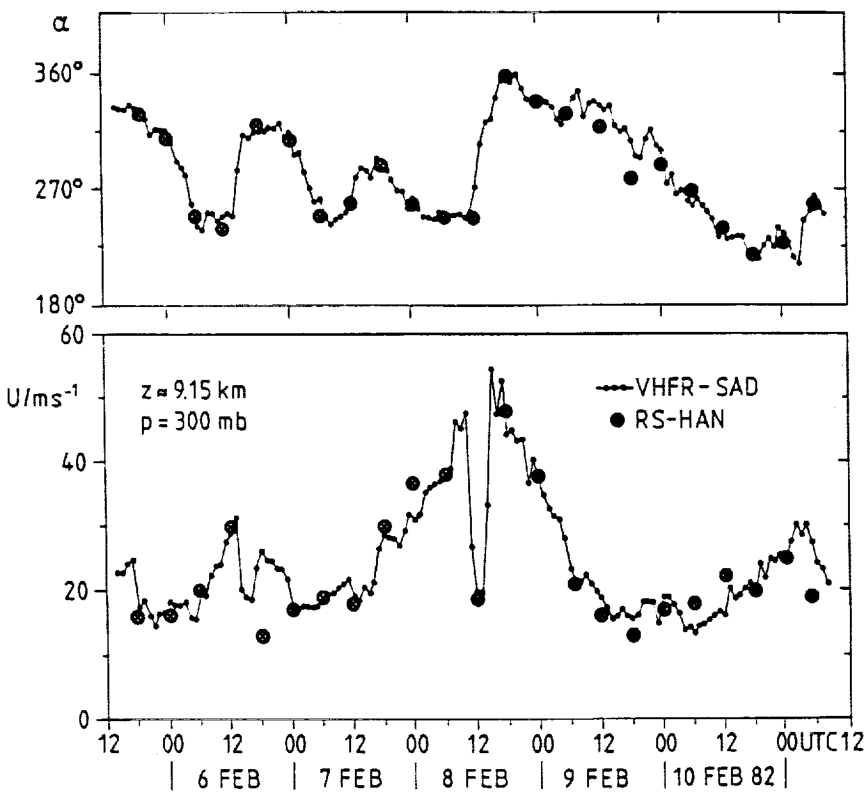


Figure 1. Time series of velocities measured with the SOUSY-VHF-Radar during February 1982 compared to the velocities measured by the nearby Hannover radiosonde at a height of 9.15 km. The solid line shows the radar measurements. The circles show the radiosonde winds.

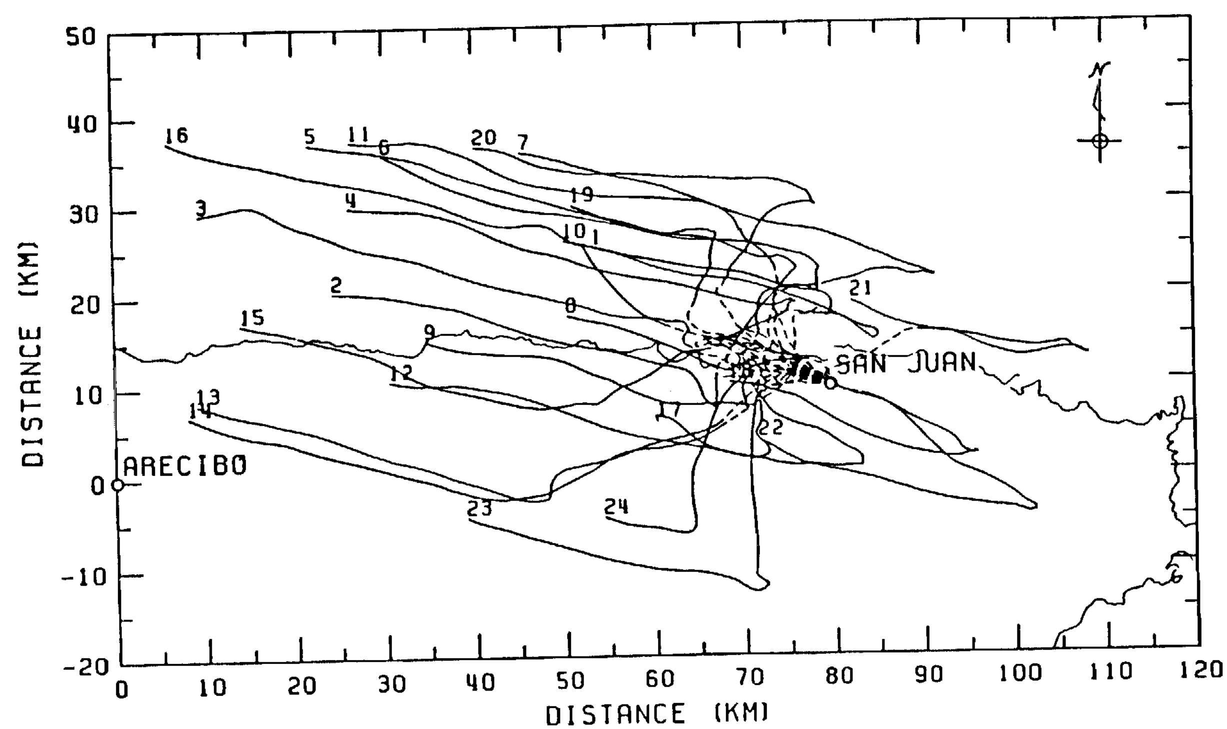


Figure 2. A map showing locations of Arecibo and San Juan, and trajectories of rawinsonde balloon ascents on 24 days in August and September 1977 when winds were also measured with the Arecibo 430-MHz radar.

spatially and temporally separated launches were carried out, and some launches from the same location at the same time were also made. The results provided an estimate of the errors attributable to the balloons and also the natural variability that occurs within the atmosphere as a function of temporal or spatial lag. Jasperson's data covered scales slightly different than the separation between San Juan and Arecibo but, when extrapolated to the appropriate scale, the natural variability could easily account for the differences between the radar winds and the radiosonde winds, without requiring large balloon errors, as pointed out by Larsen (1983).

Larsen (1983) compared data from the Poker Flat, Alaska, VHF radar to radiosonde data from nearby stations. Direct comparisons between the Poker Flat data and winds obtained with the Fairbanks radiosonde, only 30 km away, were made. Also, the radar and radiosonde winds were compared to the geostrophic winds calculated from the grid of radiosonde stations located closest to the radar. The 1-hr average radar and radiosonde winds were found to deviate from the geostrophic wind by about 5-6 m/s and they were found to deviate from each other by a smaller amount of ~2-3 m/s. Thus, the accuracy of the two wind measurement techniques appears to be comparable. The differences between the two wind measurements were again explainable on the basis of the natural variability in the flow which is a function of the spatial separation. Röttger (1983) came to a similar conclusion based on the calculated auto- and cross-correlations between SOUSY radar data and winds measured by nearby radiosonde stations.

The Wave Propagation Lab in Boulder, Colorado, has carried out a series of experiments with a 5-beam 405-MHz wind profiler to assess the precision and accuracy of the measurements first during clear air conditions over a one-month period (Strauch et al., 1987) and then during periods when precipitation was in the beam (Wuertz et al., 1988). One beam was pointed toward the vertical, and the other four beams were pointed toward north, east, west, and south at a zenith angle of 15°. The differences in the east/west and north/south line-of-sight pairs could be combined to yield independent measures of the vertical velocity. The results in the form of scatter diagrams are shown in Figure 3. Figure 3a represents (V_n-V_s) on the vertical axis and (U_e-U_w) on the horizontal axis where the subscripts refer to the direction. Strauch et al. (1987) showed that deviations along an axis running at 45° from lower left to upper right includes both the measurement error and the vertical velocity contamination of the horizontal wind measurements. The axis running from lower right to upper left includes only the measurement error. Figure 3b shows a corresponding diagram in which the vertical beam measurements have been used to correct the horizontal wind measurements. In the latter case, the scatter in the plot is more nearly circular. The measurement error was found to be 1.7 m/s without the vertical beam correction and 1.3 m/s with the vertical beam correction. The study of the

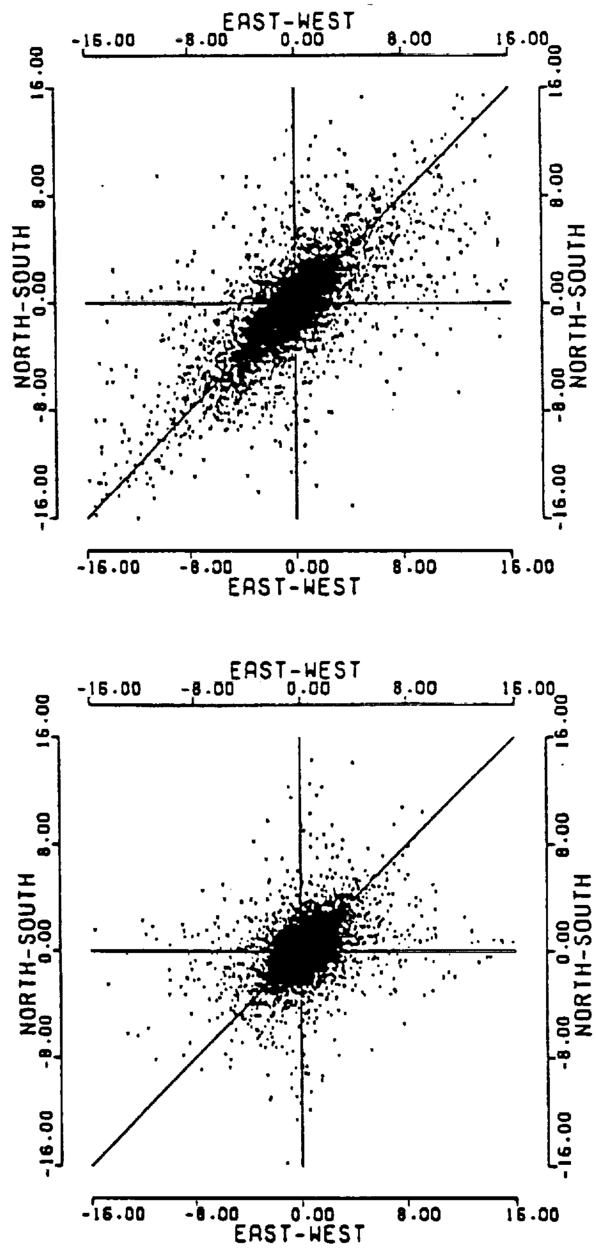


Figure 3. Scatter diagrams of [V(north)-V(south)] vs. [U(east)-U(west)] during February 1986 for the Platteville, Colorado, UHF profiler. The upper frame shows the values produced when no vertical velocity corrections were made. The lower frame shows the same data but with vertical velocity corrections applied.

precipitation cases carried out by Wuertz et al. (1988) showed that comparable measurement errors occurred when the precipitation was stratiform, i.e., spatially uniform, so that vertical velocity corrections could be made. The errors increased substantially when the precipitation was not uniform, however.

The radars not only provide better time resolution than the balloons but also overcome some other inherent limitations in the balloon technique, such as the problems that occur when balloons are launched during adverse conditions. Augustine and Zipser (1987) found that a series of balloons launched during a squall-line passage in Kansas during the Pre-STORM experiment did not reach heights much above the melting level. Wind profilers operating during the same period did not suffer from such problems. In fact, the conditions led to enhanced reflectivities that increased the signal-to-noise ratio during the period.

(b) Geostrophic Adjustment and Model Initialization

A good deal of the interest in using MST radars or wind profilers for mesoscale studies stems from the high time resolution and good temporal coverage afforded by the radars that will make it possible to study dynamical systems that have previously fallen between the "cracks" of the observational network. However, there is equally great interest in using a network of profilers as an observing system to provide the input for numerical forecast models, especially because geostrophic adjustment theory predicts that wind measurements are more useful when the scales of interest are small. The basic theory was worked out by Rossby (1937) and has been extended by numerous authors in various contexts. An excellent review is given by Blumen (1972). The theory predicts that the adjustment of the wind field and the mass field will be quite different depending on the ratio of the scale of the disturbance to the so-called Rossby radius of deformation. The Rossby radius L_R is defined as

$$L_R = \frac{C_N}{f} \tag{2}$$

where f is the Coriolis parameter and C_N is a characteristic velocity in the medium. In a shallow-water model that velocity becomes

$$C_N = \sqrt{gh} \tag{2}$$

where g is the acceleration of gravity and h is the fluid depth. Eq. 2 defines the gravity wave phase velocity in such a model. In a stratified atmosphere, the expression for C_N in a linearized treatment becomes

$$C_N^{\,2}=N^2\left(N^2c_S^{\,-2} + \Gamma^2 + m^2\right)^{-1} \tag{3}$$

where N is the Brunt-Vaisala frequency, c_S is the speed of sound, Γ is Eckart's coefficient which would vanish in the Bousinesq approximation, and m is the vertical wave number (Blumen, 1972). The Rossby radius varies substantially but typically has values in the range of 500-1000 km. The theory predicts that for scale sizes that are large compared to the Rossby radius, the wind field will tend to adjust to the pressure field. Therefore, if a synoptic scale perturbation pressure is induced in the flow, a wind that balances the pressure gradient will evolve over a time scale comparable to f^{-1} where f is the Coriolis parameter. If a large-scale perturbation in the wind field is induced, however, the energy input will be dissipated by gravity waves that propagate out of the region and carry the energy away. For small scales, the reverse situation holds, and a perturbation in the wind field will produce a pressure gradient that balances the wind geostrophically. A small-scale pressure perturbation, on the other hand, is dissipated by gravity waves that redistribute the energy.

A numerical model encompasses the same physics as the real atmosphere (or so we hope). Therefore, when observations of the pressure field at small scales are used to update the model, the effects of updated pressures are felt less. The situation is further worsened by the fact that gravity waves which are not handled well by the model are generated as part of the process of dissipating the pressure perturbation. Wind perturbations, on the other hand, tend to be maintained by the flow at small scales, and the pressure gradients needed to balance the flow develop with time. These considerations lead to an inherent preference for wind observations when small scales of motion are involved. Hoke and Anthes (1976), among others, have discussed these points in more detail, and Anthes (1983) has reviewed a number issues and strategies related to mesoscale models.

The wind profiler built and operated by NOAA's Aeronomy Lab at Christmas Island in the Pacific is the first facility to provide winds routinely by satellite transmission. After January 1987, the wind data was incorporated in the operational analysis of the National Meteorological Center and after April of that year the data was used in the European Centre for Medium-range Weather Forecasting's (ECMWF) operational analysis procedure. Gage et al. (1988) have described the results of a study designed to assess the improvement in the analysis procedure attributable to the profiler data. Before the data were incorporated in the analysis procedure, the standard deviations between the analysis and the observations was in the range of 3-5 m/s up to 250 mb. After the centers started using the profiler observations, the standard deviations decreased to 1-3 m/s. Figure 4 shows the deviations as a

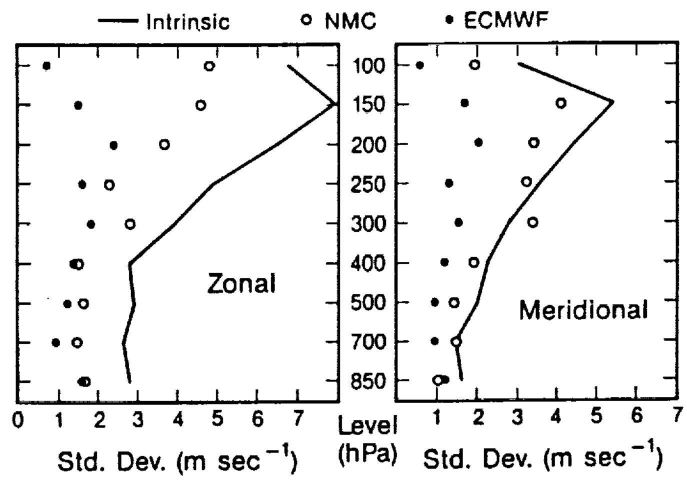


Figure 4. Standard deviations of Christmas Island winds compared to NMC and ECMWF analyses for May 1987.

function of height. The solid line indicates the variability in the profiler data alone and gives a reference by which the deviations between analysis and observations can be judged. In general, the NMC routine does a poorer job above 200-300 mb, apparently because less weight is assigned to this kind of data in carrying out the analysis. The study has shown the potential impact of wind profilers on initialization in remote areas. The minimal maintenance and unattended operation make the technique ideal for operation in isolated and remote locations. In the tropics, where the geostrophic relation is not applicable, direct observations of the winds are expected to be extremely important if significant improvements in forecasts are expected.

3. Reflectivity Data

Aside from the horizontal velocity components, the other parameter provided by all the radars is the reflectivity. The magnitude of the reflectivity is usually associated with the strength of the turbulence within the sampling volume, but strong gradients in the refractive index can also lead to enhanced reflectivities at longer wavelengths associated with what has become known as specular echoes or Fresnel reflection (see, e.g., Röttger and Larsen, 1988, or Gage, 1988). Whether the signal is due to Bragg or turbulent scatter or specular reflection, structure with a scale size comparable to the wavelength of the transmitted signal has to be present. In the former case, the structure is expected to be more microscopic in nature and characterized by shorter coherence times. In the latter case, the structure is expected to be more macroscopic in nature with longer coherence times. There is still considerable debate at meetings and in the literature about the dynamics associated with these two different kinds of scatter. Also, although we like to characterize the scatter as being of one type or the other, the actual scatter received in any real situation is likely to be the result of a combination of both effects.

These important questions need to be resolved in the future, but, in spite of the uncertainties about the details of the dynamics, we already know that in practice the result is an enhancement in the reflectivities observed at VHF at heights just below the radiosonde tropopause. Figure 5 shows an example of the reflectivities measured with the SOUSY-VHF-Radar, and the tropopause heights derived from the Hannover radiosonde data are shown by the arrows (Rastogi and Röttger, 1982). Gage and Green (1982a) have established objective criteria for locating the tropopause based on radar reflectivity data. They found that the errors were typically ± 270 m, although the authors expected the errors to increase slightly if the technique is used operationally. Gage et al. (1986) have extended the earlier work, and Sweezy and Westwater (1986) have compared different techniques for deriving the tropopause height from VHF radar data. Knowing the height of the tropopause can be valuable in

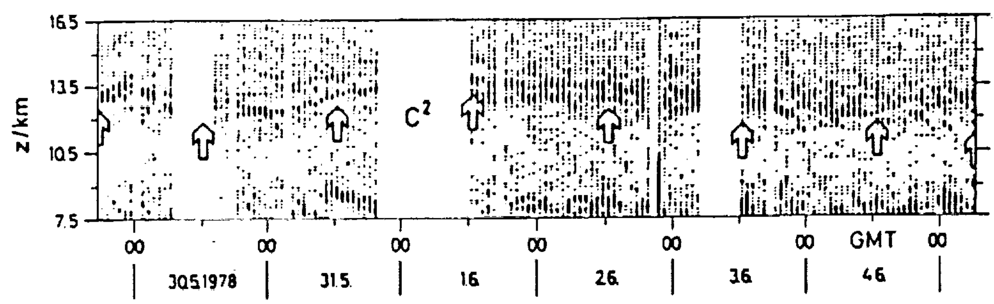


Figure 5. Plot of relative reflectivity indicating the fine structure of the tropopause region. The arrows show the tropopause height deduced from nearby radiosonde profiles.

fine-tuning the temperatures obtained from either satellite or ground-based microwave remote sensing data. Tropopause height information will not be available if UHF profilers are used since the shorter wavelengths are not sensitive to the aspect sensitivity effects that lead to the enhanced scatter from the tropopause region. Gage and Green (1982b) also pointed out that the measured radar reflectivity profiles could be used, in conjunction with information about the surface temperature, to provide an estimate of the temperature profile. Above the tropopause, the variations in reflectivitity provided information about the temperature gradients. Below the tropopause, a constant lapse rate was assumed. Radiosonde profiles will be much more detailed and accurate than the profiles derived in this way, but temperature information derived from the radar data can be used as a supplement where no other information is available. Also, the potential improvements in microwave radiometer-derived temperature profiles when the two types of data are combined have not been fully explored (Hogg et al., 1983).

An effect related to the enhanced backscatter from the tropopause seen at VHF is the enhanced scatter from frontal boundaries. Röttger (1979) was the first to describe this effect based on analysis of data obtained with the SOUSY-VHF-Radar located in the Harz Mountains in West Germany. Larsen and Röttger (1982, 1983, 1985) have analyzed a series of frontal passage events using the same instrument. In essence, the enhanced reflectivities occur very close to the location of the frontal boundary as determined from radiosonde data. The boundaries usually slope either upward with time (cold frontal passage) or downward with time (warm frontal passage). An example is shown in Figure 6 which corresponds to the passage of a cold front indicated by the cross-hatched area followed by a warm front shown by the stippled areas during the period February 7-9, 1982. The tropopause heights measured by the Hannover radiosonde are shown by the crosses. On February 8 at 12 GMT, the passage of the front caused a very low tropopause height to be reported, in agreement with the radar reflectivity contours. The lower half of the figure represents the potential refractivity M^2 calculated from the radiosonde data for the same period. M given by

$$M = -77.6\text{x}10^{-6}\left(\frac{P}{T}\right)\left(\frac{\partial \ln\theta}{\partial z}\right)\left[1+15500\,\frac{q}{T}\left[1-\frac{1}{2}\left(\frac{\partial \ln q}{\partial z}\right)\left(\frac{\partial \ln\theta}{\partial z}\right)^{-1}\right]\right] \tag{1}$$

Here P is the pressure in millibars, T is temperature in Kelvin, θ is the potential temperature, q is the specific humidity in grams per kilogram, and z is height in meters. The agreement in the gross features of both cross sections is evident. Recently, similar reflectivity structures have been observed with the Flatland radar in connection with frontal passages in Illinois (Nastrom, private communication, 1988). Larsen and Röttger (1982, 1983, 1985) made comparisons between calculated refractivities and

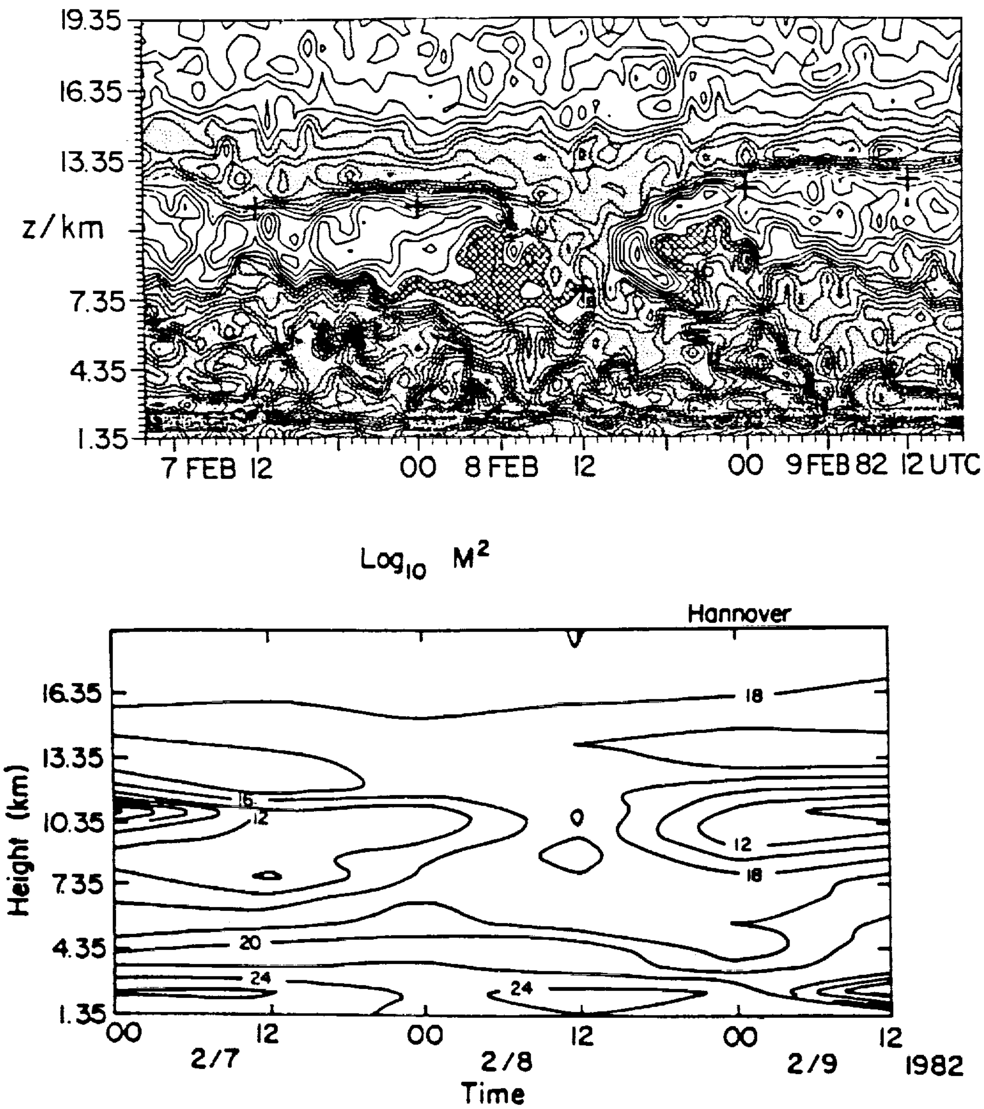


Figure 6. Top frame shows the reflectivities measured during February 1982. The stippled areas show highest reflectivities and the cross-hatched regions represent intermediate reflectivities. The lower frame shows the potential refractivity calculated from the Hannover radiosonde data for the same period.

observed reflectivities in a series of frontal passages. The agreement between the gross features of the two quantities was striking and indicated that the structure responsible for the scattering was being organized on a large scale. At least it persisted over the scale separating the radar and radiosonde which was approximately 90 km. An alternative hypothesis would have been that the enhanced reflectivities were generated by small-scale turbulence produced locally near the front. Some of the fine structure probably had this kind of source, but the larger organizing features were of mesoscale origin.

Some of the reflectivity data obtained with SOUSY has shown that the frontal boundary consists of smaller-scale filamented structure. Detailed reflectivity structure is shown in the gray-scale plot in Figure 7 for the March 6-7, 1981, warm-frontal passage. Presumably better height resolution would show even finer-scale structure. The dynamics of the upper level frontal zones are extremely important both in generating and enhancing tropospheric systems and in mixing constituents between the stratosphere and troposphere. The mixing that occurs during tropopause folding events was already documented by Danielsen (1968) in connection with experiments designed to study the transport of radioactivity and trace constituents through the tropopause region. It has become clear that the mixing between the troposphere and stratosphere is not steady and continuous but occurs discretely during certain special events, usually associated with frontal passages. We still know very little about the small-scale dynamics actually responsible for the mixing (see, e.g., Ray, 1986), but the radar reflectivities and radar measurements of velocities may provide us with more information about this aspect of the circulation.

4. Vertical Velocity Measurement Capabilities

The vertical velocity measurement capability of the MST radars has been one of the most intriguing as far as meteorological applications are concerned. Deployment of either an operational or a research network can be justified on the basis of the horizontal wind measurement capabilities alone, but vertical velocities have always had a special place in meteorology for a number of reasons. First, much of the effort in the field is devoted to predicting or understanding the vertical circulations that develop in the atmosphere. Clouds and precipitation are produced by upward vertical velocities; the large vertical gradients in density in the atmosphere lead to heating or cooling during adiabatic ascents or descents that produce significant temperature changes in competition with horizontal advection; and the vertical gradients in chemical constituent concentrations imply that vertical circulations will have a large effect on trace concentrations. In spite of the importance of the vertical velocity fields, the vertical circulation is almost always an inferred quantity. A few specialized techniques, such as aircraft measurements, exist for measuring vertical velocities, but for large-scale measurements, the vertical

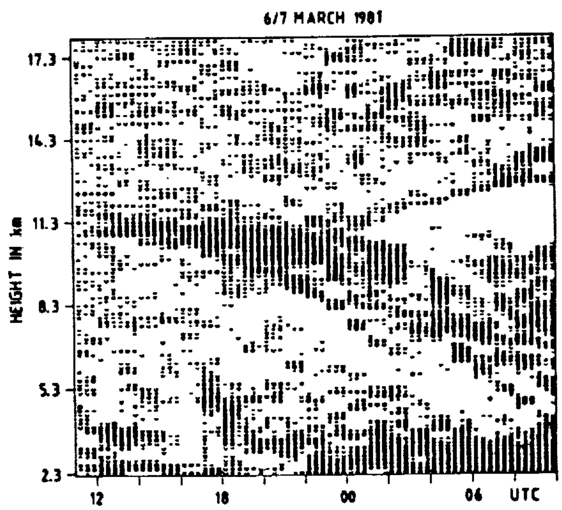


Figure 7. Reflectivities measured by the SOUSY VHF radar on March 6-7, 1981. The mean vertical reflectivity profile for the period has been subtracted from the individual profiles. Note the laminae near the tropopause and in the frontal band structure.

velocity is usually calculated from the divergence of the horizontal winds. Short-wavelength precipitation radars have been used to "measure" the vertical velocities in connection with special experiments, but the measurements usually amount to integrating the divergence over height to yield the vertical velocity. Since the radars are sensitive to precipitation, only the motion of the precipitation can be measured, and while the horizontal velocities of the precipitation and the air are expected to be the same, the vertical velocities are generally quite different.

Thus, with some reservations that I will explain later, we can say that the MST radars are the first instruments capable of providing vertical velocity measurements routinely over extended periods and over a large height range. Immediately the possibilities for verifying model forecasts of vertical velocities come to mind. Also, if the vertical velocity measurements can be trusted, the measurement of the vertical velocity can be inverted to provide us with information about the variation of divergence with height (Clark et al., 1986). The latter quantity would be extremely useful in initializing forecast models which generally use the horizontal wind field as part of the input parameters. Vertical velocity information would help the numerical models to develop realistic vertical circulations in a shorter time. Typically, the horizontal wind information is the only velocity input to the model, and it takes some time before realistic vertical circulations develop. The latter is not a problem if the ultimate goal is a 12-day forecast, but the problem is a serious one if a 2-3 hr forecast is sought. Finally, the vertical velocities may be characteristic of small-scale processes too small for the model to resolve, but even subgrid-scale momentum and heat fluxes are important in forcing the larger-scale dynamics. These fluxes are generally parameterized, but measurements of the vertical fluxes could be used to update the parameterization at the time of model initialization.

In spite of the clear need for and many potential uses of the radar vertical velocity measurements, there are still many uncertainties that need to be resolved before the potential can be realized. The first problem is related to the accuracy of the vertical velocity measurements. Although several different techniques have been used to measure horizontal velocities, including the Doppler method, the spaced antenna method, and the interferometer method, all the various techniques measure vertical velocities by the Doppler method. There seems to be general agreement that the signal received from the vertical direction is scattered by irregularities or sharp gradients in the refractive index, and that the Doppler shift of the received signal gives the line-of-sight, in this case vertical, velocity of the refractive index irregularities. The uncertainty relates to whether the irregularity velocity is the same as the air motion so that the Taylor hypothesis is valid.

A logical way to proceed in determining the accuracy of the vertical velocity measurements is to compare the measurements with vertical velocities obtained by some other technique, but in this case no real basis for comparison exists. The next best approach is to compare the measured velocities with vertical velocities calculated from other data. The number of studies of this type is very limited so far. Nastrom et al. (1985) utilized data from the Platteville, Colorado, VHF radar obtained over a period of several weeks. A number of different methods for calculating the vertical velocity were applied to the gridded values produced by the NMC analysis for the surrounding region. The kinematic method was used to integrate the divergence to produce a vertical velocity profile, the thermodynamic method was used to calculate the vertical velocity based on temperature changes associated with adiabatic warming or cooling during ascent or descent, and the quasi-geostrophic omega equation was also used. Three examples of the profiles of calculated and observed vertical velocities from the study by Nastrom et al. (1985) are shown in Figure 8. The various methods for calculating the vertical velocities, indicated by letters, were found to yield comparable results, but large differences between the measured and calculated values were found throughout the period. Specifically, the radar measurements were generally 2-3 times larger than the calculated values, but there was better agreement between the signs of the velocities.

Larsen et al. (1988) compared vertical velocities obtained with the SOUSY-VHF-Radar located in the Harz Mountains in West Germany to the vertical velocities produced by the operational analysis procedure of the European Centre for Medium-range Weather Forecasting (ECMWF). The analysis uses a normal mode initialization procedure which is known to preserve more of the divergence in the horizontal velocity field than some of the other techniques. Therefore, larger and perhaps more realistic vertical velocities are expected from this type of calculation. The result of the study was that the magnitudes of the calculated and measured values were nearly the same. However, the agreement between specific features observed in connection with the passage of fronts, for example, was not as good. An example is shown in Figure 9 which represents contours of the upward vertical velocities measured with the radar over a two-week period in November 1981 in the upper panel and the operational analysis vertical velocities for the corresponding period in the lower panel. The same gross features are present in both the observed and calculated data sets. For example, 4-5 bands of upward velocity were seen in both data sets in connection with the passage of a cold front on November 6, but the timing of the passage of the bands varied by as much as 12 hr and one band that was present in the radar data after noon on November 6 was missing in the analysis. In some sense, discrepancies of this type are not surprising, but we need to be able to account for the discrepancies in the future before we can make use of the measurements for verification and initialization of models.

Gage and Nastrom (1986) compared vertical velocities obtained with the Platteville radar located in Colorado to precipitation data from the surrounding area. They concluded that there was a

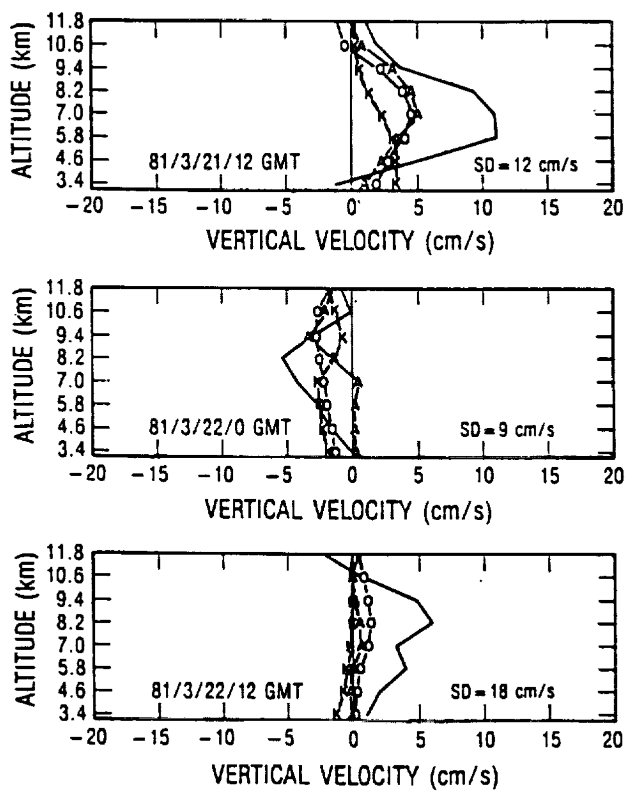
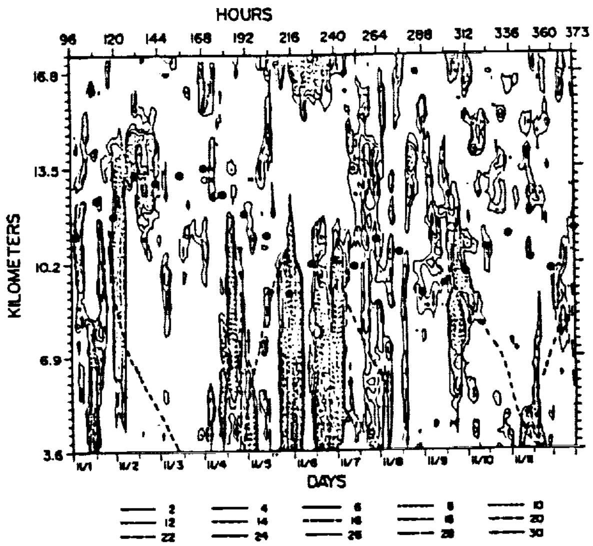


Figure 8. Profiles of the vertical velocity over Platteville, Colorado, from the ST radar (solid line) and the adiabatic (A), kinematic (K), and omega equation (O) methods.



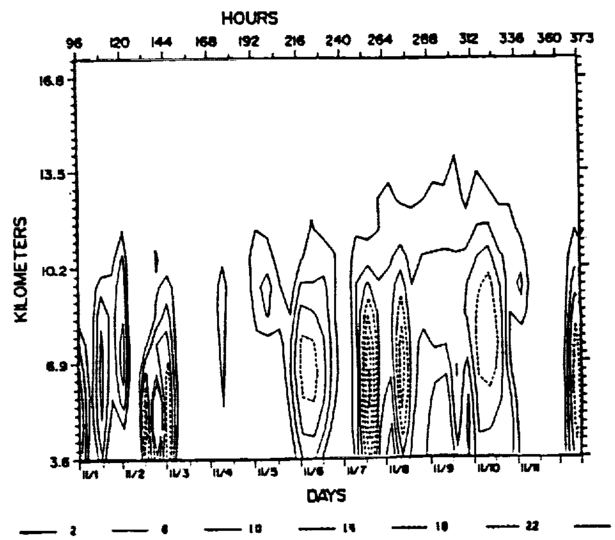


Figure 9. Contours of upward vertical velocities measured by the radar in the upper frame and produced by the ECMWF operational analysis in the lower frame for a two-week period in November 1981.

general relationship of the type that would be expected between the times when precipitation occurred downwind from the radar and the times when the radar showed a pattern of upward motion. Results of this type are encouraging, but more quantitative comparisons are difficult due to the large horizontal gradients and local variability in precipitation.

Wakasugi et al. (1985) measured the vertical velocities with the MU VHF radar during the passage of a cold front. A meteorological radar also provided data on the location of the precipitation during the period. Figure 10 is an example of the VHF data which shows the reflectivities at off-vertical incidence in the upper panel, the reflectivity at vertical incidence in the middle panel, and the circulation in the cross-frontal plane in the lower panel. The reflectivities clearly show the aspect sensitivity effects. The upward vertical velocities are strongest close to the time of the frontal passage, but a number of cells of alternating upward and downward velocities are evident before and after the passage. The times when precipitation occurred correlated well with the times when the updrafts were present, although not all updrafts produced precipitation. This type of data shows the potential for the observation of mesoscale systems and gives an indication of the insights that can be expected.

As mentioned above, an important question for the near future will be to determine the source of discrepancies between radar vertical velocity measurements and calculated values such as those produced by operational analysis. In comparing horizontal winds, a balloon can be launched immediately adjacent to the radar and a direct comparison can be made. Thus, the effects of small-scale gradients in the flow can be minimized. When the vertical velocities are calculated, usually from the divergence equation, at least three spatially separated wind profiles are needed. The result is that the derived vertical velocities represent a different quantity than the measured values since the derived values are effectively an average over the area defined by the area between the observing sites. The radar vertical velocities are representative of a single point within the domain. An attempt to reconcile the two values can be made by averaging the radar velocities in time. If the Taylor hypothesis is valid and the fluctuations are "frozen" within the flow, averaging in time should be equivalent to averaging in the spatial domain. There is some evidence that such an approach is valid (Brown and Robinson, 1979), but the two quantities will always be somewhat different. Some of these issues will likely be resolved when networks of wind profilers are deployed, such as the network proposed for the central United States as part of the STORM program (Augustine and Zipser, 1987).

Another potential problem is due to the effects of aspect sensitivity on the effective look angle of the radar beam. These effects will only be significant when longer wavelengths such as 6 m are used since the aspect sensitivity at 70 cm, for example, is negligible in virtually all but the most exceptional circumstances. Aspect sensitivity causes the largest signal to be received when the look direction is

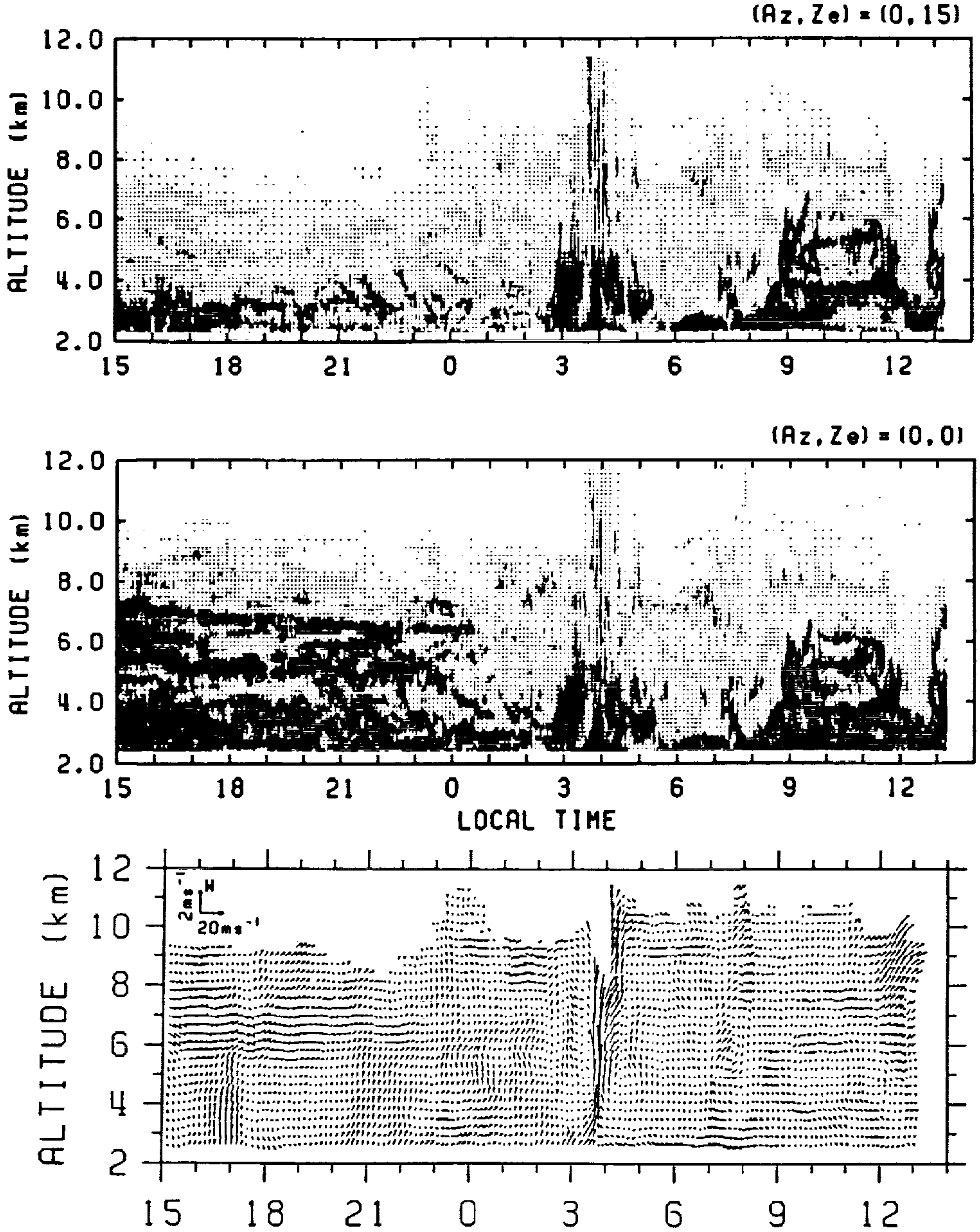


Figure 10. The upper frame shows the reflectivities measured by the MU radar in Japan at off-vertical incidence during a frontal passage in June 1984. The middle frame shows corresponding reflectivities at vertical indicidence. The lower frame shows the circulation in the cross-frontal plane during the same period.

perpendicular to strong gradients in the refractive index associated with layers in the atmosphere. The layers are generally tilted by a few degrees, but except for the largest arrays, the tilt angle will be within the beamwidth. Thus, a beam that is nominally pointed in the vertical direction may receive the strongest signal from a look direction that is a degree or a few degrees off vertical. The measurement will then consist of the projection of the true vertical velocity on the effective look direction, which is usually a minimal error, plus the component of the horizontal velocity along the look direction. The latter can create errors of over 100%, i. e., even the sign of the apparent vertical velocity can be wrong in some instances. Röttger and Ierkic (1985) have described these problems and have shown that data from a spaced antenna array can be used to calculate the tilt angle of the layers in order to make the necessary corrections to the measured velocities. Very little analysis of this type of error has been carried out so far, but the effects need to be taken into account when VHF profilers are used. Such effects may be responsible for some of the discrepancies found in the studies by Nastrom et al. (1985) and Larsen et al. (1988) since both studies used data obtained with VHF radars.

Recently a VHF radar has been constructed near Champaign, Illinois, in one of the flattest portions of the central United States. The site has tremendous advantages for studies of the vertical velocities since almost no orographic effects are expected. Figure 11 from the article by Green et al. (1988) shows samples of the time series of vertical velocities obtained at the Flatland radar, at Platteville, Colorado, just east of the Rocky Mountains, and at Sunset, Colorado, in the mountains. The variability and amplitudes of the vertical velocities increase with proximity to the mountains. The active periods that occur with some regularity in the data sets taken near the mountains are due to generation of waves by the flow over the topography.

UHF radars are unlikely to suffer from the errors introduced by aspect sensitivity effects, but the shorter wavelengths are extremely sensitive to precipitation. Thus, if there is precipitation in the beam, the measured vertical velocity will be the fall velocity of the precipitation (see, e. g., Larsen and Röttger, 1987, and Wuertz et al., 1988). Periods of strong precipitation are easily distinguished in the data since the reflectivities increase significantly beyond the levels that can be accounted for by turbulent scatter. Periods of moderate precipitation may be more difficult to handle in practice. Eliminating all questionable data is not difficult but leads to a rather unsatisfactory solution since we will be deprived of data whenever the meteorological situation becomes interesting. More work is needed to determine the optimum approach that gives us the maximum usable information but eliminates the erroneous data.

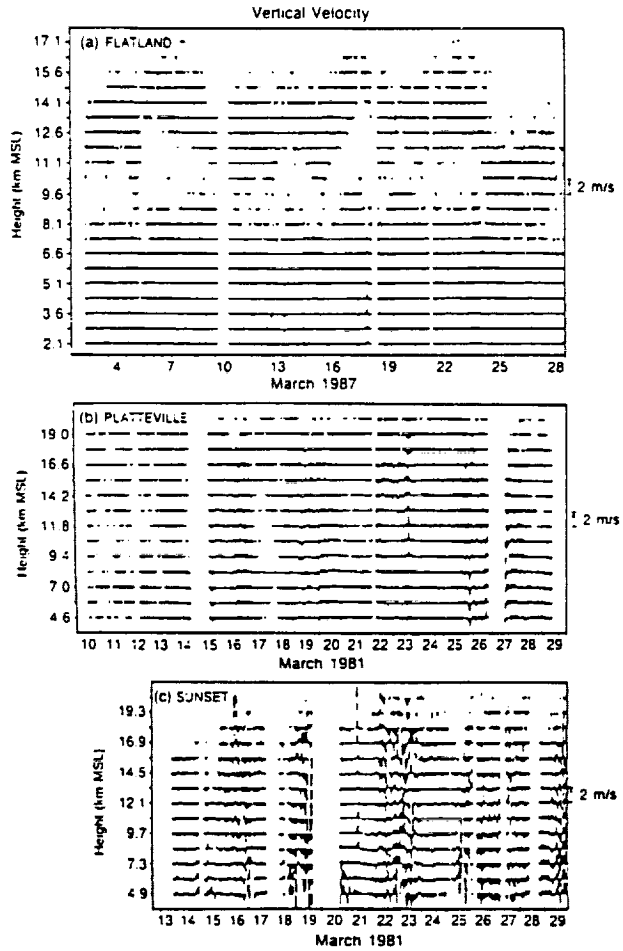


Figure 11. Radar vertical velocities averaged over fifteen-minute periods for Flatland in March 1987 in the upper frame, for Platteville during March 1981 in the middle frame, and for the Sunset radar during March 1981 in the lower frame.

5. Derivable Quantities or Parameters

(a) Pressure and Temperature Fields

The reflectivity data can be used to provide an estimate of the temperature profiles as described above, but the estimates are likely to be extremely coarse. However, there is a balance between the pressure and temperature fields and the winds imposed by the dynamical constraints. A simple example is the geostrophic relationship which relates the horizontal wind components and the horizontal pressure gradients. The balance equation is based on less limiting approximations than the geostrophic approximation and has been used in studies by Gal-Chen (1988) and by Kuo et al. (1987a,b) to determine how the wind information from a profiler network can be used to derive the temperature field. The input to the calculation is the observed wind field, and the output is a temperature field consistent with the wind field through the constraints imposed by the balance equation. The results to date have been very encouraging and show the potential for deriving parameters other than just the horizontal winds from the profiler observations.

(b) Divergence and Vorticity

Since the divergence is related to the vertical gradient of the vertical wind, the divergence can always be calculated once the vertical velocity profile is obtained (Clark et al., 1986). Alternatively, four off-vertical beams pointed along different azimuths can be combined to yield the divergence directly.

Ordinarily, a single radar cannot yield the vorticity in the flow since the radar measures the line-of-sight velocity and the vorticity is the circulation perpendicular to the look direction if the radar beam is cycled around in different azimuth directions. However, Smythe and Doviak (1987) have suggested that correlations of the signals obtained with the beams pointed in different directions can provide the cross-beam component of the flow. The procedure is similar, although not identical, to the spaced antenna method for obtaining the winds (e.g., Röttger and Vincent, 1978; Larsen and Röttger, 1988). The latter technique uses at least three vertically pointing but spatially separated beams. The cross-correlation between the signals in the various beams gives the time lag for the scatterers to move from one beam to the next. Thus, the cross-beam velocity component is derived. The spaced antenna set-up may be the most natural to use for obtaining the vorticity in the flow since the lags between the signals in the three beams are calculated anyway as part of the process of deriving the horizontal winds. The line integral of the inverse lags around the triangle defined by the receiving antennas divided by the area of the triangle will then be proportional to the vorticity.

The vorticity is an extremely important quantity in defining the properties of any meteorological flow. The vorticity associated with frontogenetic flows would be a valuable parameter to measure. Observations of this quantity could be used to study the mixing that takes place near upper-level fronts in connection with tropopause folding and in the boundary layer in connection with convection, for example. Even if this type of measurement is shown to be feasible, we still have to determine how the microscale vorticity that is measured by the radar is related to the mesoscale vorticity. The problem is similar to the one that we face in determining how representative and useful the vertical velocity measurements will be for studies of mesoscale and synoptic scale dynamics.

(c) RASS

The RASS (Radar and Acoustic Sounding System) technique was introduced already in the 1960's. The system uses an acoustic generator to produce waves with half the wavelength used by the radar. The acoustic disturbances thus produce refractive index variations that cause strong Bragg scatter of the radar wave field. Measuring the velocity of the acoustic waves yields the temperature directly since the speed of sound is a function of the temperature. The early experiments used meteorological radars with relatively short wavelengths in the centimeter range, and the high frequencies made it difficult to obtain signals much above the boundary layer. Matuura et al. (1986) were the first to implement the technique at VHF with the MU radar in Japan. The longer wavelength of 3 m for the acoustic signal decreased the attenuation and made it possible to obtain temperature profiles well into the stratosphere.

The first experiments took over a day to carry out because of the problems associated with moving the acoustic source so that the sound waves would pass through the radar beam. Since the acoustic waves are blown around by the winds, the alignment problems can be severe. Obtaining the Doppler velocity which is then converted to temperature is a relatively quick process once the acoustic beam is within the radar beam. Placing an array of acoustic sources near the radar in more recent experiments has reduced the measurement time significantly to less than one hour. The radar beam still has to be steered, but the necessary calculations and movement of the beam can be carried out quickly.

The MU radar is rather special in the sense that it is extremely flexible and powerful compared to the wind profiler systems that are being deployed operationally. Therefore, it is not immediately obvious that the RASS technique can be applied with some of the simpler systems. However, the potential uses for the temperature data that could be obtained in this way are tremendous and will, no doubt, be explored in depth in the next 5-10 years. Work has already been carried out to test the application of the RASS technique with some of the systems developed at NOAA in the Aeronomy Lab

(Ecklund, private communication, 1988) and in the Wave Propagation Lab (Strauch, private communication, 1988). The work has produced at least some modest success.

(d) Momentum fluxes

Vincent and Reid (1983) have shown that two symmetrical off-vertical beams can be used to measure the vertical flux of horizontal momentum. Other beam configurations are possible, but the Vincent-Reid set-up appears to have some distinct advantages. Only a few studies of the momentum fluxes in the troposphere and lower stratosphere have been carried out so far. Cornish and Larsen (1984) used VAD data obtained with the 430 MHz Arecibo Observatory radar, Nastrom and Green (1986) obtained momentum flux data with the Sunset radar, and Fukao et al. (1988) calculated the momentum flux from MU radar data. The latter study also compared the three beam (one vertical, two off vertical) and four beam (four off vertical) methods for obtaining the momentum flux.

The momentum fluxes are extremely important in the dynamics of the atmosphere. Gravity waves serve as the agents for redistributing energy rapidly in the vertical direction. Wave dissipation results in a convergence of the momentum flux that causes an acceleration of the flow. In particular,

$$F_v = -\frac{1}{\rho}\frac{d}{dz}(\rho v'w') \tag{10}$$

and

$$F_u = -\frac{1}{\rho}\frac{d}{dz}(\rho u'w') \tag{10}$$

where ρ is the atmospheric density, z is height, and u', v', and w' are the fluctuating zonal, meridional, and vertical velocities, respectively. F_u and F_v are the accelerations induced in the flow in the zonal and meridional directions and are usually expressed in units of $ms^{-1}day^{-1}$. Nastrom and Green (1986) found values of $\sim 10^{-2}ms^{-2}$. There was considerable variability in the data, and the direction of the momentum flux changed over intervals of 1-6 hr. Fukao et al. (1988) found accelerations of 5-25 $ms^{-1}day^{-1}$ for data averaged over a 4-day period. In the latter study, there was no evidence of large accelerations of the flow. Therefore, other processes must be balancing the induced accelerations. The study by Fukao et al. (1988) also found that the largest contribution to the momentum flux was from long period waves with periods of $\sim$300 min.

Studies of the momentum fluxes have only begun to scratch the surface of the important dynamics associated with the vertical redistribution of energy effected by the higher frequency gravity wave motions. It is likely that more studies of this type will help to clarify the important interconnections between orography and atmospheric dynamics and the mechanisms whereby energy generated in connection with dynamical processes is redistributed by the smaller-scale waves.

6. Special Operational Requirements

In certain instances, systems or networks of systems have been set up for meteorological experiments on a short-term campaign basis. However, the systems used for operational monitoring of the meteorological environment will have some special requirements in terms of strength of the system construction, reliability of the system, and accuracy.

(a) Construction and Reliability

When research systems such as the MU radar in Japan or the Flatland radar in the United States are deployed, the site is usually chosen carefully, and arrangements are made to have the site visited and maintained often. Large-scale networks generally do not afford such luxuries. Since extensive areas have to be covered by the instruments, relatively advantageous sites can be chosen, but often less than ideal choices will have to be made. Finally, the systems have to be capable of operating in a variety of weather conditions, and sometimes even in extreme weather. All of these requirements add up to the need for an exceptionally sturdy construction with a long mean time between failures.

(b) Accuracy and Precision

The required accuracy and precision are related quantities since the precision has to be at least as good as the accuracy, but the measurement error that can be tolerated is still unknown. Studies such as those of Kuo et al. (1987a, b) will help to establish the appropriate criteria. Meanwhile the systems deployed as part of the Colorado network (Strauch et al., 1987) appear to have measurement errors of about 1.5 m/s.

(c) Height Coverage

It is difficult to come up with a quantitative criteria for the height coverage required for a wind profiling system or network. In terms of the upper height limit, the more height coverage the better is probably a reasonable rule-of-thumb subject to various practical considerations. The Wave Propagation Lab in Boulder, Colorado, has compiled statistics on the height range coverage provided by systems with three different frequencies (Frisch et al., 1986). The prototype network has used a better height resolution at lower altitudes and poorer height resolution aloft, the rationale being that the larger-scale structure above still has a considerable effect on the flow closer to the surface, but the small-scale structure aloft is less likely to influence the flow at lower levels. The increased pulse length at higher altitudes increases the signal-to-noise ratio in that part of the atmosphere where the signal usually becomes weaker. At least in some cases, we know that the small-scale structure aloft is crucial in the evolution of the larger scale flow, especially near frontal zones. Then the strategy may have disadvantages.

A major concern in the meteorological community has been the lack of coverage of the lower altitudes in the planetary boundary layer rather than poorer coverage of the upper troposphere or lower stratosphere. The boundary layer is the source of important fluxes of heat and moisture that fuel the systems in the free flow above the friction layer. Therefore, measurements of winds and momentum fluxes close to the surface but above the height covered by meteorological towers are extremely important. The Aeronomy Lab and Wave Propagation Labs of NOAA in Boulder, Colorado, have built special boundary layer radars (BLR's). Ecklund et al. (1988) have described one of the systems. These instruments use high frequencies that allow fast transmit/receive switches to be utilized. Since the signals in the boundary layer are relatively strong, small antenna sizes can be used. Already, these small specialized systems have shown good results, and the systems appear to provide good overlap with the larger profilers.

(d) Choice of Frequency

So far, only a few frequencies have been used in profiler applications. The popular choices to date have been near 50 MHz and 400 MHz. The ideal frequency, even for a particular application, is not known because the range of the frequency spectrum that has been explored is so limited. In practice, it may be almost impossible to choose any frequency but the 405 MHz frequency that is already allocated for meteorological use. The advantages of the 405 MHz band are that large bandwidths are easier to obtain, making it possible to operate with better height resolution, and faster T-R switches are generally available at the higher frequencies so that better coverage of the lower heights can be realized. The disadvantages include lack of aspect sensitivity, which eliminates the enhanced scatter from the tropopause or frontal boundaries, and the sensitivity to precipitation. Even low rainfall rates will result in the signal being dominated by the precipitation so that direct measurements of the vertical air motion become unreliable or impossible.

7. Remaining Questions

(a) Best Configuration for Networks

The prototype profiler network being deployed as part of the STORM program in the United

States is being established with an average spacing comparable to that of the radiosonde network, although a subgrid array with a smaller spacing will be embedded in the larger array. The plan seems reasonable, but it is still unclear whether this is an optimum spacing. We will need experience with the network to determine what characteristics of the various dynamical systems can be observed and whether important features are slipping between the cracks.

(b) Initialization of Models

There are still many important unanswered questions about how the profiler data can best be used to initialize numerical forecast models. Clearly, the straightforward input of profiler winds at the initial time step is a likely possibility, but various four-dimensional data assimilation schemes may also be useful, especially since the wind profilers have tremendous time resolution but provide limited spatial coverage.

(c) Usefulness of Vertical Velocity Measurements

We have already described some of the potential problems with the vertical velocity measurements with regard to the effect of small-scale variability on the representativeness of the measurements. In addition, we need to determine if the measurements can be used in model initializations in some way. Verifying model output with the vertical velocities is possible, but it may be that the implied divergence information can be used to generate more realistic vertical velocities at an earlier stage in the model run. The profile of vertical velocity can also be indicative of the latent heating associated with convection. Such information would be useful in parameterizing the heating and momentum fluxes generated by clouds. The errors introduced by the different types of scatter and by the effects of precipitation need to be determined more accurately in the future so that we can make the best use of the vertical velocity information.

(d) Usefulness of Small-Scale Turbulence Parameter Measurements

There is still disagreement about the dependence of the reflectivity and spectral width on the turbulent dissipation rate, although the work of Hocking (1983) has served to clarify a number of these issues. These small-scale fluxes can be extremely important in certain dynamical situations and can be potentially important in improving parameterizations of subgrid scale motions in numerical models. However, more comparisons between radar-derived values and observed quantities are needed to gain confidence in the turbulence data that can be obtained from the radar measurements. Also, Hocking et al. (1986) have shown that the spectral width measurements will be strongly contaminated by beam broadening and shears for any typical flows within the troposphere. We need to determine if other radar techniques can alleviate this problem and provide comparable information in a different way.

(e) Usefulness of Momentum Flux

The momentum fluxes measured by the Vincent-Reid (1983) technique appear to have a great potential for improving the parameterizations of small-scale dynamics used in virtually all of the forecast models. Again, the representativeness of fluxes measured at a single location will have to be determined. The possibility of updating the parameterizations based on flux-measurement input at initialization time should be explored.

(f) Advantages of Various Radar Techniques

So far, all the operational profiler applications have used the Doppler technique which involves two or more beams pointed in off-vertical directions. The combination of the line-of-sight velocities measured in the various beams produces the wind components. Alternative methods for measuring the winds include the spaced antenna method and the interferometer technique. The spaced antenna method has already been tested sufficiently so that we know that it can be used routinely. There are even various advantages of the technique for certain applications. The interferometer technique has not been tested extensively and its operational potential is unknown, but it provides some possibilities for measuring small scale structure in the flow that is impossible with other techniques, at least in the simplified implementations typical of the operational profilers.

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