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MESOSCALE METEOROLOGY

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Introduction

The last four years have seen a considerable expansion in research on mesoscale atmospheric phenomena. The motivation is three-fold: the prospect of greater forecast accuracy, an emerging ability to observe the mesoscale, and the challenge of understanding extremely complex physical mechanisms. The current level of accuracy of storm warnings and other local forecasts can be substantially advanced only by an improved ability to deal with structures smaller than the traditional synoptic scale. These structures are beginning to be described by analysis of special data and of routine surface observations. Nonlinear dynamics, often combined with nontrivial cloud microphysics, poses formidable theoretical problems. Hence, the research and operational communities have recently formulated initial plans for a national project on mesoscale meteorology (UCAR, 1982).

The definition of mesoscale phenomena is not universally accepted; the simplest one is morphological: "mesoscale" refers to those systems which are too large to be observed completely from a single point (lacking a capability for remote sensing) and too small to be observed unambiguously by the routine upper-level sounding network over the continental areas, with station spacing of a few hundred kilometers (Ligda, 1951). A physical definition, on the other hand, seems better

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as a guide to research. It has been suggested that mesoscale circulations are dynamically characterized by the importance of the Coriolis force, but not by the dominance that assures quasi-geostrophic flow. This definition appears to exclude higher-frequency gravity waves and all equatorial disturbances except those of low frequency. It would include intense extratropical cyclones, which have been regarded as synoptic scale events. Some other physical characteristic, such as the relative importance of cloud microphysics, might also be used as a definition, but many mesoscale phenomena do not involve condensation. Although an entirely adequate physical definition may not be necessary, a dynamically self-consistent definition of "mesoscale" seems desirable. For the present, however, we shall err on the side of insufficient, rather than excessive, exclusivity. We will not, however, aim for completeness, except in the bibliography. In this text we will discuss only a few lines of research that we find especially interesting and important.

We will distinguish between free circulations and topographically induced ones. The former arise from some sort of instability in the larger-scale atmospheric structure, while the latter represent the direct response to processes due to features of the earth's surface and are more or less anchored to them.

Among the free circulations we find organized convective systems, and structures within cyclones of synoptic scale, encompassing the fronts of extratropical latitudes and the cyclonic rainbands both in and outside the tropics. At the larger end of the mesoscale range, we will discuss the small and sometimes intense maritime cyclones of middle latitudes. At the smaller end of the mesoscale range, recent work on generation and propagation of gravity waves, inertia-gravity waves, and internal bores will be presented.

The influences of topography which produce the forced circulations may be thermal, arising from differences between land and water or in type of land surface and vegetative cover, from varying slopes of mountainous terrain, or from prominent gradients in sea-surface temperature. Alternatively, the circulations may be orographically produced. Recent work on these various phenomena will be described.

In closing, we discuss two important subjects not easily classified on a phenomenological basis: the transfer of energy through mesoscales, in which paradoxical results are being obtained which demand physical explication, and regional scale numerical modeling, which attempts to predict the development of mesoscale circulations within initially smoother synoptic-scale flow.

Organized Convective Systems

During the quadrennium just completed, more research has been devoted to organized cumulus convection than to any other mesoscale phenomenon. Much of this work has been inspired by the results of special field programs including those conducted in Oklahoma by the National Severe Storms Laboratory (NSSL), in the eastern tropical Atlantic during the GARP Atlantic Tropical Experiment (GATE), and a number of others. Much other work was based on the close study of data available routinely in the United States, typically for cases of spectacular phenomenology.

From these observations, a picture has emerged which seems to be valid throughout a large range of latitudes, confirming suggestions and inferences from earlier studies by Newton (1950) and Fujita and Brown (1958) for middle latitudes, and by Zipser (1969) for the tropics. Initially, there is sporadic convection (e.g. Leary and Houze, 1979b; Maddox, 1980a); then the convection consolidates and a thick layer of stratiform cloud appears in the middle and upper troposphere, producing long-lasting and significant amounts of rain and having in some respects a life of its own. This layer represents the accumulation of debris from the convection, which remains active at its upshear edge, and may be aided by additional condensation in an internal mesoscale region of ascent. Finally, after perhaps twelve hours, the active convection becomes disorganized and loses its identity.

The ubiquity of this picture notwithstanding, it is useful to distinguish between the squall system and the system recently entitled the Mesoscale Convective Complex (MCC) (Maddox, 1980a). The former propagates rapidly over the surface and with respect to the airflow in most of the troposphere (e.g. Fortune, 1980; Gamache and Houze, 1982; Ogura and Liou, 1980). The MCC moves more deliberately, often with the mean tropospheric flow (Bosart and Sanders, 1981). Leary (1979) found, in an example from GATE, weak cyclonic circulation in the lower and middle troposphere, while Bosart and Sanders pointed to the resemblance of their overland mid-latitude example to a weak tropical storm above a shallow and erratic boundary layer. In all cases there is pronounced anticyclonic circulation in the upper troposphere.

The initiation of these systems is still something of an observational puzzle. In fact, there are too many possibilities. The notion of interaction of jets in the lower and upper troposphere originated with Beebe and Bates (1955) and has been recently advanced by Uccellini and Johnson (1979). Circumstantial evidence points to the "dry line" of the Central Plains, separating warm humid air to the east from hot dry air to the west, as the locus of storm initiation (e.g. Burgess and Davies-Jones, 1979; McCarthy and Koch, 1982), but the mechanism is still in question. In Florida the peculiar character of the topography produces a characteristic interaction between the sea-breeze and convective systems (Cooper et al., 1982). Local hot spots, arising from peculiarities of the rugged terrain (Cotton et al., 1982), trigger convection in Colorado and no doubt elsewhere. Gravity waves interact with convective systems, characteristically being produced by them, but sometimes (Uccellini, 1975) initiating them. Strong low-level warm advection sometimes appears to be responsible (Maddox and Doswell, 1982). Symmetric instability (Ogura et al., 1982; Emanuel 1979, 1982) is rapidly gaining popularity, while horizontal variations in the diurnal development of the surface boundary layer appear to be important over land (Colby, 1980; Garrett, 1982; Ogura et al., 1982; Sun and Ogura, 1979), and ongoing convection, especially over the tropical oceans, appears to provide a continual stimulus for development of new systems. Randall and Huffman (1980), in a sort of reductio ad absurdum, have provided theoretical evidence that cloud-induced stabilization and destabilization, if reasonably distributed with respect to the convective element, can produce convective clusters by a completely stochastic process.

It is likely that all of these possibilities are valid under the proper conditions. A dose of eclecticism, however, is sorely needed.

Rawinsondes and conventional radar are still the main data sources for observational studies of organized convective systems, but special surface mesonetworks (e.g. Chagnon, 1981; Cotton et al., 1982; McCarthy and Koch, 1982) and satellite information are finding increasing application. Satellite cloud imagery is used by Mack and Wylie (1982) to infer upward tropospheric mass flux from the rate of expansion of cirrus anvil near the tropopause. Clouds, as tracked by geosynchronous satellite, provide information on wind (e.g. Halpern, 1979; Johnson and Suchman, 1980), while results presented by Hillger and Vonder Haar (1979) and Chesters et al. (1982) indicate an ability to map, from infrared radiance data, the horizontal structure of the surface boundary layer with useful accuracy, at least so far as temperature in concerned. The successful sensing of moisture, with this as with all techniques, continues to be elusive.

Compared to observational studies, theoretical research of mesoscale properties of convection has proceeded at a modest pace. Most of the theoretical work conducted during the period focuses on four subjects: 1) The structure and dynamics of tropical convective systems, with emphasis on the mesoscale updrafts and downdrafts found in the anvil region to the rear of the active convective clouds; 2) The nature of Mesoscale Convective Complexes (MCC's); 3) Basic properties of truncated spectral models of moist convection, and 4) The triggering mechanisms for frontal and pre-frontal (middle-latitude) squall lines.

The enormous complexity of moist convective processes makes it difficult to identify the underlying causes of mesoscale organization of moist convection using analytical theoretical approaches. This has led most investigators to use numerical models as tools for ultimately understanding these processes. The success of this approach depends upon the proper incorporation of the relevent physical processes in the models as well as a thoughtful and complete diagnosis of the model results.

A particularly imaginative application of numerical models for studying MCC's was introduced by Maddox et al. (1981). Rather than engage in the difficult task of representing moist convection in a mesoscale numerical model, the authors compare the results of a model run without a convective representation to the actual behavior of an atmosphere which contains significant convection; the differences are attributed to the convection. Averaging the results of a number of cases involving MCC's lends additional credence to their findings, which agree reasonably well with detailed observational studies such as those of Bosart and Sanders (1981). The picture which emerges is that of a large convectively driven thermal cyclone, with strong anti-cyclonic circulation in the upper troposphere and weak cyclonic flow just above the evaporatively cooled boundary layer. These features are also evident in purely numerical simulations of MCC's such as those of Fritsch and Chappell (1980a,b). The identification of the life cycle, structure, and circulation associated with these storms is an important achievement. Much however, remains to be learned about the dynamical interaction between moist convection and the mesoscale circulations which are so clearly evident in MCC's.

Among the more intriguing observational findings (discussed earlier in this section) is the existence of mesoscale regions of low-level subsidence, high-level ascent, and stratiform precipitation to the rear of most tropical and some middle-latitude squall lines (e.g., Gamache and Houze, 1982; Ogura and Liou, 1980). Recent theoretical research has been directed toward an explanation of the downdrafts (Brown, 1979; Leary, 1980), and the incorporation of the mesoscale circulations in budget analyses and convective representations (Houze and Cheng, 1981). Both Brown and Leary find that the magnitude of the mesoscale downdrafts can be explained by evaporative cooling, while Brown (1979) finds additionally that the qualitative nature of the behavior of the convective system depends on the presence of the evaporative cooling associated with the mesoscale region of precipitation. Houze and Cheng (1981) use one-dimensional models of convective and mesoscale drafts, constrained to conform to budget estimates based on GATE data, to evaluate the effect of mesoscale drafts on budgets of heat, moisture, and mass. They conclude that the mesoscale updrafts and downdrafts lead to an increase of the ensemble average mass flux at upper levels and a decrease at lower levels, while the heat fluxes of the mesoscale up- and downdrafts tend to nearly cancel, leaving estimates of the ensemble average based on cumulus fluxes alone essentially unchanged. One of the important theoretical problems remaining is the cause of the mesoscale updraft to the rear of the active convection.

The dynamics of middle-latitude squall lines in strongly baroclinic flows has remained an important subject of theoretical research. Emanuel (1979) proposed that convection occurring in shear-parallel rows would preferentially occur in regions of low or negative symmetric stability, and showed that the latter was an essentially mesoscale instability, occurring at moderate Rossby number. Ogura et al. (1982) showed that symmetric instability might have initiated a squall line which occurred during the Severe Environmental Storms and Mesoscale Experiment (SESAME) in 1979. The addition of parametric moist convection to a symmetric baroclinic flow was shown by Emanuel (1982) to result in distinctly mesoscale circulations which propagate to the right of the mean shear and in some other respects resemble middle-latitude squall lines. Sun and Ogura (1979) demonstrated that low-level convergence, associated with differential mixed layer development in the presence of initial horizontal surface temperature gradients, is capable of initiating squall-line convection, and proposed that this process was effective in initiating the severe convection on 8 June 1966. Squall lines associated with frontal systems were numerically simulated by Chang et al. (1981) and by Ross and Orlanski (1982). The former used a convective representation while convection was produced explicitly by the latter. Both were moderately successful in simulating some of the observed features of frontal squall lines. An intriguing aspect of the simulation of Ross and Orlanski is the decoupling of the surface fields of divergence and vorticity during the course of the development; this decoupling was attributed to a change in the propagation characteristics of the convection and is remarkably similar to patterns seen in the analytical solutions of Emanuel (1982).

A fruitful analytic approach to the investigation of the characteristics of moist convection involves the use of highly truncated spectral models which retain a modest number of nonlinear wave-wave interactions. This type of model was first applied by Lorenz (1963) to the study of Bénard convection and has recently been extended to the moist case by Shirer and Dutton (1979), Shirer (1980), and Yost and Shirer (1982). These studies show that in mildly supercritical conditions a large variety of unstable modes is possible, including both propagating and steady convection. A branching hierarchy of modes was established and the mechanisms whereby some modes lose stability to others were investigated. The effects of mean vertical shear and horizontal temperature gradients were also explored. One of the more interesting results of Yost and Shirer (1982) is that horizontal heating gradients have a singular effect on the structure of Bénard modes, so that the modes obtained in a model without such horizontal gradients are unobservable. Although the implications of such models for atmospheric convection are not altogether clear, it appears that much can be learned from simple analytical models of this kind.

Some additional studies, though not essentially directed towards mesoscale phenomena, deal with cumulus convection and thus seem appropriate for brief mention here. Betts (1982a,b) formulated a new way of discussing moist-convective processes, using as a central concept the properties of an air parcel, cloudy or not, at the point where it is marginally saturated with respect to water vapor. The methodology is readily adaptable to graphical representation, but its clear advantage over traditional methods becomes apparent only when a quantitative measure of the non-vapor water content of cloudy air is available, though this measurement, of course, is not routinely made.

Some observations made from a sailplane in a Colorado cumulus, presented by Paluch (1979), indicate that entrainment occurs mainly through the tops of the clouds, as originally suggested by Squires (1958), rather than through the sides, as has been generally held. Emanuel (1981) studied theoretically the conditions in which downdrafts from the cloud top could penetrate deeply to become effective mixing agents. He found that, with appropriate vertical profiles of cloud water and temperature, this mixing could be highly effective, and he suggested that penetrative downdrafts could be responsible for mamma aloft and for small-scale downbursts of strongly divergent wind near the ground. The consequences of this view of cumulus-entrainment for estimating the bulk effect of convection in circulation systems of larger scale are not yet clear.

Severe Thunderstorms

Because of its great destructiveness in the United States (Fujita, 1973; Staff, NSSFC, 1980) and because of the fascination it holds for the intrepid naturalist (Bluestein and Sohl, 1979), the severe thunderstorm (accompanied, as a matter of definition, by damaging hail, or wind, often tornadic) has recieved much attention in the past four years. Substantial progress was made in research on both observational and theoretical aspects of this phenomenon.

The distinct character of the severe thunderstorm was foreshadowed by Brooks (1949), who described a larger cyclonic circulation within which certain prominent tornadoes were embedded. The availability of radar allowed inferences to be drawn concerning the three-dimensional structure of these violent and relatively persistent convective storms, denoted "supercells" by Browning (1964).

Recent Doppler radar analyses (e.g., Ray et al., 1980, 1981) confirm the earlier speculative view. In short, the supercell thunderstorm resembles a very much scaled-down but very intense baroclinic cyclone with a major buoyant updraft which rotates cyclonically in the lower troposphere and then diverges to leave the storm as a thick anvil-like cloud aloft in an anticyclonically curving trajectory, principally in the direction of the large-scale vertical wind shear vector. Beneath this cloud is the "foreward-flank downdraft", driven by evaporative cooling from the heavy precipitation. The cool air spreads out at the surface and impinges on the warm moist air in the path of the storm, not unlike a warm front in an ordinary cyclone, although the source of the cold air is different.

At the rear of the storm, the overtaking upper-tropospheric flow separates, as if the main updraft represented a solid obstacle, while part of the mid-tropospheric flow sinks upon contact with the storm, again as a consequence of evaporative cooling, to become the "rear-flank downdraft." This downdraft rotates cyclonically around the storm in the lower troposphere, with a cold-front-like "rear-flank gust-front" at its leading edge. The associated "flanking line" of growing cumulus towers has long been known to harbor tornadoes on occasion, but the major tornadic events tend to occur in the region of high vorticity near the base of the rotating updraft, especially (Lemon and Doswell, 1979; Brandes, 1981; Klemp and Rotunno, 1982) in the narrow zone of strong contrast between the warm inflowing updraft air and the surrounding rain-cooled air from the downdrafts.

Not infrequently (e.g. Klemp et al., 1981) this cold air wraps around the updraft in an occlusion-like process which destroys it near the ground but encourages a new cyclonic updraft center at the subsequent "point of occlusion." The perseverance of the updraft aloft is attributable to the rapid conveyance of condensate to the surrounding downdraft regions. Thus excessive water loading is avoided and, unlike the situation in the synoptic-scale cyclone, the major precipitation occurs with descending, rather than ascending, motion. The powerful updraft itself is typically a region of weak radar reflectivity, as the condensed water particles are dispersed by the upper-level divergence before they can grow to radar-detectable size.

Considerable effort has been devoted to the application of visible and infrared satellite imagery to identification of severe thunderstorms. For example, Adler and Fenn (1979a, 1979b) found that the outflow divergence and cloud-top ascent rates were approximately twice as large for severe thunderstorms as for lesser ones. High-plains hailstorms were found by Reynolds (1980) to display cloud-top temperatures substantially and persistently colder than the ambient tropopause. McCann (1981) and Negri (1982) identified severe thunderstorms on the basis of a V-shaped region of minimum cloud-top temperature at the upwind edge of the storm, with a limited embedded downwind area of relative warmth. A tendency for tornadoes to occur just after strong cloud-top ascent, often during a period of shrinkage of cold cloud top and of radar reflectivity, was identified by Adler and Fenn (1981) and by Wexler and Blackmer (1982). In a more general approach, detailed mapping of low cloud-motion vectors in selected cases showed a coincidence between subsequent severe thunderstorm development and preceding mid-tropospheric vertical motion on the 100-km scale (Wilson and Houghton, 1979), or preceding moisture convergence on the 40-km scale (Negri and Vonder Haar, 1980).

The forecasting of severe thunderstorms still proceeds largely along the lines laid out by Fawbush et al. (1951). That is, for a prominent outbreak of major tornadoes and other severe manifestations there must be 1) substantial potential instability in a narrow wedge of warm moist air in the lowest km or so, 2) intersection of (or impingement upon) this wedge by a jet of strong winds in the middle troposphere, and 3) a triggering mechanism. Forecasting skill, however, remains very small (Murphy and Winkler, 1982), and the correct physical interpretation of these requirements is not entirely clear.

Although the source of energy for severe thunderstorms is probably largely the same as that which drives less intense varieties of moist convection, there appear to be differences in both the degree of instability present and the manner in which it is released. There are few systematic studies of these differences. A promising though limited set of results presented recently by Weisman and Klemp (1982) suggests, from numerical simulations, that the buoyancy available to the updraft must be appropriately matched to the large-scale vertical wind shear.

The role of the strong winds aloft may be a direct one, for Weisman and Klemp find that substantial shear is necessary for the production of a persistent rotating storm resembling the supercell. Alternatively, the role may be an indirect one, since the strong winds may harbor mobile short waves with the capability for destabilizing the air column by cooling aloft; or the strong vertical and lateral shears associated with upper-level jet structure may provide a base state which is unstable with respect to mesoscale perturbations, which can trigger the severe convection itself. Mesoscale perturbations, not necessarily of this character, as reflected in the surface pressure field, were found by Miller and Sanders (1980) to modulate the convection by producing an increase in the number of severe events in the spectacular case of 3 April 1974.

The suggested triggering mechanisms are largely those regarded as appropriate for less violent convective systems. The proximity of strong shallow thermal contrast, however, appears to be peculiarly favorable for tornadic storms, as implied originally by Darkow et al. (1958) and as found in more recent case studies by Zipser and Golden (1979), Hoxit et al. (1980) and Maddox et al. (1980).

The significance of shallow thermal boundaries may be their association with strong vertical wind shears. The horizontal vorticity represented by this shear appears, through the tilting of vortex tubes, to be the source of the vertical vorticity of the supercell mesocyclone, as indicated in modeling studies by Schlesinger (1980), Wilhelmson and Klemp (1981), and Klemp and Rotunno (1982), and also as inferred from observations by Brown and Knupp (1980).

Given the growing understanding of how to view the structure of the ambient atmosphere and an increasing ability to predict diurnal development of the boundary layer, and assuming an improvement in ability to monitor mesoscale variability from in situ surface observations or from remote sensing, the forecasting of supercell thunderstorms with attendant severe phenomena should improve markedly in the next few years, with respect to both lead time and spatial resolution. The production of a significant fraction of damaging wind and hail events by less characteristically recognizable convective systems, however, may limit this forecasting capability to relatively modest levels.

Mesoscale Structures Within Extratropical Cyclones

The first widely studied mesoscale structures, although they were not so denoted, were the fronts embedded in extratropical cyclones. Discovered by the Bergen school, they were assigned what is now seen as an excessively exclusive role in cyclogenesis and the production of rain. Still, they remain a fascinating feature which limits the accuracy of local forecasting.

The first case study explicitly labelled as mesoscale was an investigation of convection along a cold front by Swingle and Rosenberg (1953). Here, radar and surface hourly observations were combined to show that the precipitation structure was organized, although not as simply as in a single solid line along the frontal discontinuity. The multiply banded structure of frontal precipitation has been exhaustively described by the CYCLES project, for both cold fronts (Hobbs et al., 1980; Herzegh and Hobbs, 1981) and warm fronts (Herzegh and Hobbs, 1980; Houze et al., 1981). Studying systems near the coast of the Pacific Northwest with Doppler radar, aircraft, serial balloon soundings and surface mesonetworks, they found in virtually all cases that precipitation at the ground was greatly enhanced by the superposition of arrays of "seeder" cells in the middle and upper troposphere, above "feeder" bands of stratocumulus below. Mesoscale vertical motions ranged from a few tenths of 1 m s^{-1} to about 1.5 m s^{-1}.

The dynamical cause of the feeder bands was for the most part not addressed by the CYCLES group, although Hobbs and Persson (1982) suggest that the wave-like corrugations in a cold-front rain band arise either from a dynamic instability due to lateral shear of the wind or from a gravitational instability due to frontal overhang near the surface.

In a study of a spectacular case in the Central Valley of California, Carbone (1982) analyzed triple-Doppler observations showing a cold-frontal updraft of 15-20 m s^{-1} over a limited area. Tornadoes accompanied this front, despite the lack of convective buoyancy, representing a dramatic departure from the usual production mechanism. Evidently, strong ambient wind shear can compensate, in some way not yet clarified, for lack of potential instability in the thermal and moisture stratification. In this case the front as a whole moved like a gravity current, and it appeared that melting of precipitation may have represented a significant thermodynamic forcing for the dynamical behavior of the system.

Another possibility, which is related to the aforementioned presence of strong wind shear, is that shear-parallel frontal rainbands are manifestations of symmetric instability, as first proposed by Bennetts and Hoskins (1979) and Emanuel (1979). Although the conditions for this instability are rarely satisfied in the unsaturated atmosphere, the possibility of phase changes of water may render the flow conditionally unstable to symmetric instability. The conditions for moist symmetric instability are far more easily satisfied in the atmosphere, and the resulting circulations have much in common with observed rainbands, including their alignment with the vertical shear and their mesoscale dimensions. Emanuel (1982) has shown that the conditions for the onset of conditional symmetric instability as well as certain aspects of the structure of the resulting circulations may be assessed using circulation integrals.

In addition to investigations of the numerous banded structures observed within extratropical cyclones, considerable research has been directed toward a refinement of the understanding of surface frontal dynamics, together with a continuing effort to explain the causes of fronts found in the middle and upper troposphere. The latter are especially interesting in their association with strong turbulence events in the free atmosphere, and in their possible role in exchanging mass with the stratosphere. Shapiro (1980) has estimated that turbulence associated with upper-level fronts and folds in the tropopause is one of the primary means by which chemical constituents are transported through the tropopause, while Gidel and Shapiro (1979) indicate that such turbulence provides significant sources and sinks of potential vorticity. In addition to these processes, turbulence may also drive secondary circulations in the vicinity of upper-level fronts (Shapiro, 1981).

The mechanism of frontogenesis near the tropopause remains something of a mystery. While early analytical models such as those of Hoskins and Bretherton (1972) lead to a form of high-level frontogenesis, the general features of the modelled fronts show significant differences from the classical observational work of Reed and Sanders (1953). Strictly two-dimensional models have difficulty in reproducing the strong thermally indirect circulations frequently observed in the vicinity of upper-level fronts, since conversion from kinetic to potential energy is not possible in symmetric models unless the Ertel potential vorticity is negative and thus the flow is symmetrically unstable. Strong horizontal temperature gradients resulting from thermally indirect circulations may occur more easily in flows which are not strictly two-dimensional. Shapiro (1981) has shown that the inclusion of the action of shearing deformation upon an along-front temperature gradient may lead to significant subsidence warming in the frontal region; this appears to be in better agreement with the observations of Reed and Sanders (1953). Recent three-dimensional semi-geostrophic modeling by Heckley and Hoskins (1982) shows the importance of deceleration of the flow downstream from the upper-level ridge.

Current research on surface fronts may be described as a continual refinement of and improvement upon the seminal numerical work of Williams (1972) and the analytical nonlinear models of Hoskins and Bretherton (1972), with main emphasis on the inclusion of planetary boundary-layer processes and effects due to latent heating. Blumen (1980) performed a detailed comparison of the results of the Hoskins-Bretherton model with the analysis of an intense surface front by Sanders (1955), and observed that while the general aspect of the modelled and observed fronts were similar, certain small-scale features were significantly different. The main problem of the models was their failure to simulate the narrow vertical jet which Sanders observed just ahead of the surface front. The models' performance in this regard was somewhat improved by the incorporation of an Ekman boundary layer by Blumen (1980), and was more dramatically improved by the inclusion of more complete boundary-layer physics by Keyser and Anthes (1982). It appears that while the general dynamics of surface fronts are described elegantly by simple models based on the geostrophic momentum approximation (Eliassen, 1948; Hoskins and Bretherton, 1972), the nature of the frontal circulations close to the front itself and within and just above the boundary layer require more detailed treatments for their explanation.

The incorporation of conditional latent heating in models of frontogenesis renders the analytical approaches intractable, and requires careful consideration in formulating numerical models. Williams et al. (1981) included grid-scale condensational heating in their numerical model, and performed dry and moist adiabatic adjustments in order to account for sub-grid-scale heating due to convection. Although these parameterizations are somewhat crude, they were probably adequate, since the simulations were performed for circumstances in which convection was weak and of secondary importance. Williams et al. were able to demonstrate that the condensational heating strengthens both the horizontal temperature gradients and the frontal circulations in the region above the planetary boundary layer, and that for sufficiently strong heating, smaller-scale circulations imbedded in the frontal zone occur. These resemble the symmetric instabilities found by Bennetts and Hoskins (1979) in a numerical model, which have also been suggested by Emanuel (1979).

The simulation of frontal circulations in the presence of significant conditional instability is a yet more difficult task. Ross and Orlanski (1982) attempted to numerically simulate an observed front which was associated with moist convection in the form of intermittent squall lines. Rather than relying upon parametric convection, the authors simulated the convection explicitly, albeit coarsely. They were able to demonstrate that substantial changes in the frontal circulations result from moist-convective processes, and were able to simulate with reasonable accuracy several of the observed features of the frontal weather. In addition, and as has been noted previously, the frontal convection during the latter part of the simulation showed signs of decoupling from the front itself, as is often observed.

In recent years, it has become apparent that boundary layer processes may produce shallow fronts which, however, sometimes have a noticeable effect on clouds and precipitation. The New England coastal front (e.g. Bosart et al., 1972) is a good example of such a front, and is often associated with a local enhancement of precipitation within large-scale cyclonic storms (Marks and Austin, 1979). Many mechanisms have been proposed to explain the existence of the coastal front, which forms generally during disturbed winter weather, with on-shore geostrophic winds. The two which have received the most attention are horizontal gradients of boundary layer heating, due to heat transfer from the ocean surface, and convergence forced by differential surface roughness. Undoubtedly, the classical frontogenesis mechanism operates in the late stages of frontal collapse. A comprehensive numerical model, which includes boundary layer processes and terrain, was described by Ballentine (1980), who used the model to simulate coastal frontogenesis starting from observed initial and boundary conditions. His results appear to suggest that heat transfer from the sea surface and latent heating are the main forcing mechanisms, while friction plays a secondary role. The issue, however, is by no means settled.

Another interesting example of boundary-layer processes operating to produce frontal circulations was described by Bluestein (1982). In this case, a small (mesoscale) area of cold air near the ground was produced locally by diabatic cooling over snow-covered terrain. The resulting circulation appeared to be responsible for an unpredicted region of clouds and low temperatures in Oklahoma. As better observations of boundary-layer processes become available, it seems likely that more mesoscale phenomena, such as those discussed here, will become apparent.

Small-Scale Cyclones

Two not entirely exclusive examples of small cyclones, with horizontal scales of 100-1500 km, have received recent attention; both are predominantly maritime phenomena of the extratropical cold season.

The first example (Reed, 1979; Mullen, 1979) occurs not on a pre-existing surface front, but rather in the broad baroclinic zone on its poleward side. This "polar low" has its own cloud system, distinct from the main frontal cloud band seen in the satellite imagery. Sometimes, but by no means always (Locatelli et al., 1982), there is a sympathetic wave development at a point on the main front adjacent to the polar low. Almost always there is an eventual connection in the cloud structure between the low and the major frontal band, having the appearance of a classical occlusion, although arising from a process different from that envisioned by the Norwegian school. Some polar lows (e.g. Rasmussen, 1981) form deep in the trough of cold air, with only slight vertical wind shear and baroclinic forcing.

There is some controversy concerning the physical character of polar lows. Reed (1979) and Mullen (1979) show that the structure of their Pacific examples is clearly that of a baroclinic cyclone, and Locatelli et al. (1982) report associated rainbands similar to those seen by the CYCLES Project in larger cyclones. Rasmussen (1981) argues for the significance of convection, showing a warm core for his sub-Arctic Atlantic case. All studies show substantial conditional instability in the lower troposphere, and Mullen (1979) notes vigorous heat transfer from the sea surface. No doubt both parties are right, and the disagreement stems from differences in the choice of cases studied.

The second example is an explosively developing intense inner core of a larger cyclone, denoted a meteorological "bomb" by Sanders and Gyakum (1980). These dangerous storms tend to develop near regions of strong sea-surface temperature gradient in the western portions of both the Atlantic and Pacific Oceans, although counterexamples can be found. This cyclone is clearly baroclinic, developing (shallowly at first) in advance of a mobile trough in the middle troposphere, which has a previous history of travel across the upwind continent. "Bombs" develop either along the warm edge of the major large-scale baroclinic zone, like a classical wave cyclone, or within or toward the cold edge of the zone, like a polar low.

This class of storms shares with polar lows substantial conditional instability in the lower half of the troposphere, and satellite imagery provides clear indication of associated deep convection, on the southern and eastern periphery of the storm rather than over the center, although an eye-like structure is sometimes seen (e.g. Bosart, 1981).

Current numerical models (Sanders and Gyakum, 1980) do not adequately predict this explosive cyclogenesis, partly because of limited horizontal and perhaps vertical resolution, but also partly because of inadequate reckoning of the effects of convection and of transfers of heat and moisture to and within the surface boundary layer. In a linearized quasi-geostrophic model, Mak (1982) finds that the addition of heating, related to vertical motion in lower levels as a way of simulating convective effects, increases growth rate and decreases the preferred wavelength by factors of five and three, respectively, relative to the result without such heating. Satyamurty et al. (1982) confirm the importance of heating for small-scale cyclogenesis, in a theoretical study of a primitive-equation model.

Mesoscale Wave Generation and Propagation

The advent of high-power Doppler radar in recent years has led to an increasing number of observations of small-scale and mesoscale wave events in the atmosphere. Previously, such waves could be mapped only by closely spaced arrays of microbarographs; and could be detected to a limited extent by aircraft, rockets, and rawinsondes. None of these conventional methods allows assessment of the three-dimensional nature of the wave propagation.

Gravity waves are relatively easy to detect at high altitudes, since the upward decrease of ambient density requires the velocity amplitude of the waves to increase, following the upward progress of the wave group. High-frequency FW-CW pulsed Doppler radars have been used to detect gravity waves in the F-region of the ionosphere. Hung et al. (1980) used ray tracing to determine that an isolated thunderstorm was the source of ionospheric gravity waves with periods of around 15 minutes. Similar results had been found by Hung et al. (1979). Gravity waves excited by convection can be an important energy source for the mesosphere and ionosphere. Clark and Morone (1981) found that intense heating of the mesosphere, measured by rockets launched from Wallops Island, Virginia, accompanied the passage of thunderstorms and attribute the heating to the dissipation of gravity waves excited by the convection. These waves may constitute a particularly large energy source for the upper atmosphere in summer, when convection is most active and when upward propagation of planetary waves is inhibited by easterlies in the stratosphere. To date, it appears that most of the gravity-wave activity in the ionosphere is related to tropospheric thunderstorms and jet streams, and to various instabilities which occur in the auroral region.

Doppler VHF radar has also been used to detect gravity waves in the troposphere. Ecklund et al. (1981) have documented fluctuations in the vertical winds at two sites, above and 60 km east, respectively, of the front range of the Colorado Rockies. These fluctuations are most likely due to flow over the mountains, since they are markedly weaker at the downwind site and since their intensity is well correlated with the zonal component of the 500-mb flow. Ecklund et al. (1981) also find that gravity-wave events may be associated with intense baroclinic zones, characterized by strong vertical wind shear. A study by Van Zandt et al. (1979) presents evidence that tropospheric gravity waves can be generated by dynamical instabilities of strong jet streams.

Other means by which inertia-gravity waves may be generated have been recently proposed by Ley and Peltier (1978 and 1981) and Chimonas et al. (1980). During the latter stages of frontal development, the geostrophic adjustment of the front-parallel velocity component may occur rather rapidly. Ley and Peltier (1978) have suggested that gravity waves may be produced during such an adjustment; these waves may then serve to initiate pre-frontal squall lines. A dramatic example of wave propagation in the warm sector of a baroclinic cyclone occurred in connection with the severe tornado outbreak of 3-4 April, 1974, as documented by Miller and Sanders (1980). Here, some of the waves appeared to precede the convection and may have been initiated by frontogenetical processes, although no strong front was in evidence at the surface. Definitive documentation of wave generation by frontal processes is still lacking.

Gravity wave generation by convection is still a controversial and interesting subject of observational and theoretical research. A recent example can be found in the study of Balachandran (1980), who detected gravity-wave activity within the cold-air outflow from thunderstorms. While there can be little doubt that inertia-gravity waves are produced by moist convection, the ability of the wave and the convection to react synergistically, as suggested by Yamasaki (1969), Lindzen (1974), and others, remains open to question. A difficulty encountered in the theoretical formulation of this problem is that the time scales of the wave and the convection are not sufficiently different so that conventional cumulus representations, which rely on scale separation, can be applied straightforwardly. Indeed, the application of such representations in the case of inertia-gravity waves invariably leads to a monotonic increase of growth rate as the wavenumber increases, with no short-wave cutoff. Emanuel (1982) shows that a short-wave cutoff does occur when inertia-gravity waves are driven by convection in a strongly baroclinic environment; such waves have a maximum growth rate at horizontal scales which make the characteristic Rossby number about unity. Chimonas et al. (1980) circumvent the scale-separation problem by including condensation directly in the linear wave equations; these are then meant to show only the early influence of non-convective condensation on wave growth. Their results suggest that when condensation occurs near the wave's critical level, strong amplification of the wave may result. It appears that the familiar criterion for wave-over-reflection at critical levels, namely that the Richardson Number be less than 1/4 there, may be satisfied in a more stable atmosphere provided that condensation reduces the effective Richardson Number sufficiently.

The internal bore, an interesting phenomenon which has been studied extensively in Australia (e.g. Clarke et al., 1981), has recently been documented in the United States by Shreffler and Binkowski (1981). As is apparently the case with the Australian "Morning Glory", the bore appears to propagate at the top of a stable nocturnal boundary layer surmounted by a deep layer of nearly neutral stratification. The bore studied by Shreffler and Binkowski propagated through a large portion of the midwest at speeds of about 50 km hr^{-1}, and was accompanied by pressure rises of 1-2 mb. It seemed to originate in a region of strong thunderstorms, and its passage was apparently accompanied by the turbulent collapse of the nocturnal low-level jet. Collapse has also been documented in connection with the presence of crashing bores. The prevalence of these phenomena in the atmosphere, and the conditions under which they may occur, are not well understood and constitute important subjects of research.

Thermally Forced Circulations

The land-sea breeze, and the mountain-valley wind, two phenomena whose study must have begun with the origin of meteorology itself, continued to attract a modest level of attention during the past four years. As to the first, Estoque (1981) presented a detailed observation study of the development of the lake breeze over Lake Ontario on an October day with light southerly geostrophic wind. Starting at the west end of the lake at 1000 LST, the breeze spread eastward, achieving complete coverage by 1900 LST. The circulation veered with time in response to the Coriolis force, and extended only to 450 m. A contrasting behavior in Equatorial latitudes on the Brazilian coast was studied by Sun and Orlanski (1981). In this instance a series of inland cloud bands parallel to the coast, spaced a few hundred km apart, supported the idea of propagating waves to be expected as the Coriolis force becomes small. Sun and Orlanski's linear stability analysis indicated that the response to the sea breeze could be viewed as a trapeze instability in this instance.

Particular local regions display particular characteristics. The sea-breezes on the east and west coasts of Florida are known to interact strongly with convection over the peninsula. Burpee (1979) showed that this sea-breeze convergence is weaker in the surface winds on days with heavy convective rain than on days without, as the divergence of evaporatively cooled sub-cloud air dominates the entire area by late afternoon. On the Oregon coast, a modeling study by Clancy et al. (1979) showed an interaction between the sea breeze and coastal upwelling in the ocean. The feedback was weak, however, because of the relatively small scale of the atmospheric perturbation.

Great Lakes snowstorms, from shallow clouds unconnected with cyclone-scale ascent and associated cloudiness, received increasing attention. Passarelli and Braham (1981) found from results of a special field program that the land-breeze convergence into Lake Michigan was a prominent aspect of such storms there, while Kelly (1982) observed roll structure in the clouds, with axes parallel to the low-level wind. In a mesoscale modeling study, Ellenton and Danard (1979) showed that transfers of moisture and heat from the lake were the primary agents of heavy snowfall near the lake shores, with orographic lifting and frictional effects playing secondary roles.

In a field study of mountain and valley circulations in South Park, Colorado, Banta and Cotton (1982) found the usual nightime drainage wind and a shallow morning upslope breeze. By afternoon, however, deepening of the convective boundary layer resulted in the appearance of westerlies at the surface, as a consequence of the downward mixing of upper-level momentum. In a modeling study of downslope winds, Manins and Sawford (1979) found that entrainment of such air at the top of the katabatic layer exerted a stronger influence on the character of the flow than did the surface stress. Heating over the San Mateo mountains of New Mexico was probed by aircraft, as reported by Raymond and Wilkening (1980). They found a toroidal circulation of the "heat-island" type above a peak.

Both orographic and land-sea thermal effects are evident in a study by Mass (1982), of the diurnal circulations of western Washington. The effects are much stronger during summer, when the synoptic-scale forcing is weaker. The diurnal variations of precipitation and of the horizontal divergence in the surface winds are related, but in a complex and indirect way.

Urban heat-island effects in St. Louis during summer METROMEX were elaborated by Shreffler (1979), who found a greater convergence over the city during the day than at night. There was some evidence of a daytime influence on convective storminess.

Fog, surely thermally produced and perhaps associated with mesoscale flow patterns, received some attention. Noonkester (1979) studied field data obtained in the Pacific near the coast of southern California, finding that radiative cooling at cloud top was an important effect. Variations of fog-top height often appeared to be controlled by mesoscale circulations. Mesoscale ebb-tidal circulations in the ocean appeared, in a study by Woodcock (1982), to produce a minimum in sea-surface temperature at the western end of the Cape Cod canal. This coldness, acting upon a mixture of two nearly saturated air masses of different temperatures, was found to be the proximate cause of the fog which preferentially occurs there.

Orographic Circulations

The influence of topography on atmospheric flow has continued to be an important subject of research. Flow over mountains induces a variety of wave motions varying in scale from planetary Rossby waves to non-hydrostatic gravity waves. The former have been the subject of considerable research, particularly since it appears that nonlinear interactions between the mean flow and

Rossby waves may be involved in the important phenomenon of atmospheric blocking. The smaller-scale motions have also received a great deal of attention, in part because they are associated with many local weather phenomena, including severe downslope windstorms, mesoscale vortices which are sometimes destructive, and a wide variety of precipitation events including flash floods. Here we focus attention on the smaller-scale motions, which we define so as to exclude Rossby waves and to apply to flows for which the Rossby number is no more than unity. Here the Rossby number is defined as

$$R_o = \frac{U_o}{fa} \tag{1}$$

where U_o is a characteristic flow speed, f is the Coriolis parameter (assumed constant), and "a" is a characteristic horizontal scale of the topography.

Historically, most theoretical research on orographic gravity waves dealt with simple flows for which planetary rotation (f) is negligible, and nonlinear effects are small. Most studies are also confined to the consideration of flows over two-dimensional ridges. The past decade has, however, seen a rapid extension of the earlier work to include strongly nonlinear effects, rotation, and effects due to three-dimensional topography. Observations of orographic flows have also increased in precision with the advent of portable automated surface stations, Doppler radar, satellite-based systems, and research aircraft equipped with inertial navigation systems. Flow over complex topography has been a particularly active subject of observational studies in recent years.

The latter is perhaps best exemplified by the work of Reed (1980), who investigated the causes of the destruction of the Hood Canal Bridge in western Washington in early 1979. Detailed examination of surface wind observations, including those taken at the site of the bridge, reveal the presence of an intense mesoscale cyclonic vortex which evidently formed as an effect of, and in the lee of, the Olympic Mountains. Similar flows have been discussed by Smith (1982a), who also develops a linear theory for flow past a three-dimensional isolated mountain, valid for large Rossby number but including some effects of the earth's rotation. The theory indeed predicts low pressure in the wake of the mountain; this pressure disturbance is hydrostatically related to subsidence warming in the lee of the mountain. While the wind field does show cyclonic rotation in the vicinity of the pressure low, the wind is far from being geostrophically balanced and tends to flow through the pressure disturbance. A certain mystery remains, however, in the case of the Hood Canal Bridge. Smith (1981) argues persuasively that since the effective Rossby number in this case is roughly ten, rotation should not be evident in the perturbed flow, but the observations (Reed 1980, 1981) leave little doubt as to the presence of a closed circulation.

Under different synoptic conditions, flow past the Olympic Mountains apparently produces a somewhat more rectilinear zone of convergence in the lee of the mountains (Mass, 1981). This zone is often associated with an enhancement of clouds and precipitation and resembles in other ways a shallow cold frontal circulation. The conditions under which this kind of flow may be induced by mountains have not yet been accounted for theoretically.

Under conditions in which gaps or passes through a range of mountains are roughly aligned with the flow, or where channels cut through relatively high topography, high winds through the gaps sometimes occur. When flow over orography is stratified, it acts somewhat like two-dimensional potential flow (Ito, 1982) and is driven by the pressure differential across the mountain which is in turn related to damming of cold air to the windward and subsidence warming in the lee of the mountains (Reed, 1981). Gap winds, some of which resemble Boras, have been recently observed in the sounds of northern Baffin Bay (Ito, 1982), the Strait of Juan de Fuca in western Washington (Overland and Walter, 1981), and various passes through the mountains of western Washington (Reed, 1981).

An interesting phenomenon which sometimes accompanies flow over two-dimensional ridges is the rotor, which takes the form of a closed overturning cell just downwind of the ridge. This circulation results in local upslope flow at the surface just under the rotor. While such winds are generally weak, a case in which they caused substantial damage was documented by Zipser and Bedard (1982). It was not possible to find anything in the large-scale flow to distinguish this case from the more commonly observed severe downslope windstorm.

Baroclinic effects on flow over orography have just begun to receive some attention. Smith (1982b) notes that the blocking effect of mountains on the low-level flow may lead to the overturning of moist or dry isentropic surfaces when cold advection is occurring, and proposes that the resulting convection may constitute an important facet of orographically enhanced rainfall. When the Rossby number of the flow is not too large, a mountain-parallel jet may be formed as a consequence of the cross-mountain temperature gradient which arises due to adiabatic temperature changes. Parish (1982) demonstrates, using a numerical model, that such effects can account for the low-level jet sometimes observed on the windward slopes of the Sierra Nevada.

The effect of planetary rotation on orographic flows when the Rossby number is not small enough to allow the application of quasi-geostrophy is the subject of papers by Smith (1979, 1982a). In the former, Smith shows that the mountain-induced linear wave drag for stratified rotating flow over a two-dimensional ridge decreases with decreasing Rossby number; this decrease is associated with a diminished upward angular momentum flux. In the second paper, the second-order effects of rotation (large Rossby number) on flow over a three-dimensional bell-shaped mountain are examined by expanding the linear, inviscid, Boussinesq equations in the small parameter $1/R_o$. The zero-order solutions are obtained by extending the classical results for a two-dimensional bell-shaped mountain of Queney (1948) to three dimensions (Smith, 1980). Near the surface, this flow exhibits lateral and vertical deflection, with a pressure ridge to windward and a trough to leeward of the mountain. To first order, planetary rotation affects only the horizontal velocity components near the ground, leaving the pressure and vertical velocity fields unaltered. The Coriolis-induced horizontal deflections are as expected, with ageostrophic flow down the pressure gradient on the windward side of the mountain where blocking and flow deceleration occur. These deflections result in a slowly-decaying train of inertia waves (with frequency f and wavelength $2\pi U_o/f$) in the lee of the mountains, as earlier described by Queney (1948). These waves have not been observed, perhaps due to the absence of an associated pressure signal.

In addition to planetary rotation, the presence of cross-stream variation of orography has certain important effects on the flow. It has long been known that an isolated hill will excite both transverse modes (with wavefronts perpendicular to the flow) and diverging diagonal modes, which together have been called "ship waves" (e.g. Wurtele, 1957) since they resemble the familiar pattern of external modes which occur behind objects moving on the sea surface. Recently, fully three-dimensional linear models of flow around isolated mountains were advanced by Smith (1980), Simard and Peltier (1982), and Blumen and Dietze (1981). Smith (1980) obtained the linear solutions for constant flow past an idealized bell-shaped three-dimensional mountain, while Simard and Peltier (1982) constructed a model which allowed for arbitrary vertical variations of flow and stability, and thus were able to compare their solutions to flow observed in the lee of isolated islands. Blumen and Dietze (1981) included a cross-stream flow variation (but no vertical variation) in a linear model of flow past isolated hills of various shapes, and found that a reasonable cross-stream variation, localized near the mountain, effectively confined the wave motions to a narrow strip extending downwind from the mountains. In other respects, their solutions resemble those of Smith (1980).

An effect which is not seen in linear models of flow past isolated mountains is the tendency for the mountain to shed vortices, particularly when the stratification is strong. These vortices result from the instability of horizontal shears created by flow around the obstacle. That this phenomenon is common in the atmosphere is dramatically illustrated in satellite photographs of the flow of a stable layer around isolated islands, resulting in a well defined Karman vortex street made visible by stratocumuli. Recently, a band of westerly winds extending 150 km downwind of Hawaii in an otherwise easterly flow was documented by Nickerson and Dias (1981).

The effect of nonlinearity on flow over two-dimensional mountains also continues to be an interesting and somewhat controversial subject. The nonlinearity of purely orographic flows is measured by the aspect ratio of the topography, H/a, where a and H are characteristic horizontal and vertical scales, respectively. Long (1972) was able to show that the linear steady solutions which obtain for hydrostatic flow over a bell-shaped mountain, when the upstream flow and stability are constant, are also solutions of the nonlinear equations, provided that the fully nonlinear lower boundary condition is applied. This model predicts, inter alia, that when the aspect ratio exceeds a critical value (about .85 U/Na, where U is the flow speed and N the Brunt-Väisälä frequency), the wave streamlines become vertical at certain altitudes and a local critical level occurs, resulting in greatly increased wave-induced drag. Peltier and Clark (1979) showed that when the waves actually break, the resulting wave-induced turbulent layer in which the static stability is effectively neutral serves to reflect subsequent waves. If this level occurs at an altitude representing an integral number of half-wavelengths in the vertical, then, as pointed out by Klemp and Lilly (1975), constructive interference will occur between the incident and reflected wave, resulting in increased wave drag and strong surface winds. While Peltier and Clark argue that a non-hydrostatic model is necessary to produce the convective overturning which leads to this effect, Lilly and Klemp (1980) point out that hydrostatic models can produce turbulent layers through shearing instability, which they argue may be more important than convective instability. The possibility of self-induced critical layers is of great interest in the general field of internal wave dynamics, and perhaps such regions play an important role in severe downslope windstorms.

Mesoscale Turbulence and Energy Transfer

In addition to providing a wealth of information on mesoscale waves, Doppler radar and aircraft observation have led to a rapid advance in the understanding of the behavior of turbulence on the mesoscale. Analysis of energy spectra obtained through detailed observations of atmospheric flows can help identify the sources and sinks of atmospheric energy, and can reveal certain aspects of the nature of processes which transfer energy from the scales at which it is generated to the scales at which dissipation occurs. For example, the existence of a spectrum in which the spectral energy density depends on the minus three power of the wavenumber, K, is a strong indication that the motions are two-dimensional in character, being so constrained at the large scale by rotation, and perhaps at smaller scales by stratification; while a minus five-thirds power law is indicative of three-dimensional motions. The nature of the spectrum also indicates whether energy is being transferred to larger or to smaller scales.

The most revealing and exciting discovery in this topic in recent years has been the determination that energy spectra seem to obey the minus five-thirds law for wavelengths less than about 1000 km. Evidence for this relation was reviewed by Gage (1979), while more advanced measurements using Doppler VHF radar have been presented by Larsen et al. (1982). Classical similarity theory for three-dimensional turbulence in the inertial subrange, where sources and sinks are not considered to have direct effects, suggests that energy should cascade to smaller scales through a -5/3 power law, but it is difficult to argue that this behavior should characterize atmospheric motions at scales much larger than about 100 m. At the other end of the scale, Charney (1971) and others argued that geostrophic (two-dimensional) turbulence should cascade enstrophy downscale in such a way that the energy spectrum should range as the minus third power of the wavenumber; this power law is in good agreement with observations of large-scale motions.

Gage (1979) proposed that the -5/3 power law observed for mesoscale motions represents two-dimensional turbulence in an inertial range, which transfers energy to larger scales. It was also suggested that the observed mesoscale energy might be due to a spectrum of internal gravity waves (Dewan, 1979). Lilly (1983) analyzed some of the properties of turbulent wake collapse in stratified fluids and concluded that Gage's proposal that mesoscale energy may result from an upscale energy transfer by two-dimensional turbulence may well be correct; the source of such energy is likely to be convection and shear instability. Only a small percentage of energy generated by these small-scale processes needs to cascade upscale through two-dimensional turbulence to explain the observed spectra; the rest is, presumably, lost to three-dimensional turbulence and gravity waves.

As further observations of turbulence on the mesoscale become available, it seems likely that the nature of sources and sinks in the mesoscale domain will be better understood, as well as the character of those processes which act to transfer energy between scales. If, indeed, energy is transported upscale at the large end of mesoscale and enstrophy is transported downscale through the synoptic scales, then energy and enstrophy sinks are implied at intermediate scales. To the authors' knowledge, such sinks have not been identified. These questions constitute an important basis for further research.

Regional Scale Numerical Modeling

More or less conventional numerical modeling using relatively small horizontal mesh lengths (50-100 km) and with relatively complex representations of convection and of surface boundary-layer processes is often regarded as part of mesoscale research. Some related examples, with rather finer meshes, were described in the foregoing sections. Here we discuss the so-called regional-scale modeling, in which the goal is to predict not the mesoscale system itself, but rather the environment in which it develops. This favorable environment may exist over an area of the order of 200 km across, and hence requires a model with relatively high horizontal resolution for its proper elucidation; some operational models approach this resolution.

Regional-scale numerical models were developed at a number of locations, including Pennsylvania State University and Drexel University. Recent results from the former were directed towards predicting the environment of severe thunderstorms. Forecasts described by Carlson et al. (1980) and by Anthes et al. (1982) indicate that fronts, jets, and thermodynamic features favorable for intense convection can be predicted 24 or more hours ahead, beginning with relatively bland initial conditions. These predictions employed SESAME rawinsonde data.

The Drexel model, reported by Chang et al. (1981), was used with routine data to forecast a case which produced a line of severe convection, including a major tornado at Omaha. The narrow zone of potential instability, produced by differential thermal advection, became even narrower as the horizontal mesh length was reduced from 140 km to 35 km.

It appears, then, that on at least some important occasions a number of the conditions for severe convection can be predicted with a precision limited only by the available computer resources. Nor is the usefulness of fine-mesh numerical prediction limited to this kind of situation. Daily experience with operational models indicates frequent success with a variety of relatively small-scale events in wind, temperature, and precipitation. Failures, however, are not uncommon.

Concluding Remarks

It is clear that mesoscale research is a vigorous and growing enterprise with active and challenging naturalistic, experimental, and theoretical aspects. Of these three we perceive theory to be the least advanced, a situation hardly new to meteorology. Except with respect to frontogenesis and perhaps the land-sea breeze, where relatively advanced understanding has been achieved, satisfactory theoretical elucidation is still lacking. In particular, the physical mechanism of the organized mesoscale convective system eludes us; we know many detailed characteristics but lack the knowledge of why they appear as they do. The pervasive bandedness of precipitation challenges our tenuous grasp of the theory of mesoscale instabilities, especially where phase changes of water play a key role. We appreciate the pervasiveness of gravity waves, but do not understand clearly their interaction with other mesoscale entities which are perhaps more interesting to us and which have a more practical impact.

We would like to see increased interest in two topics. One of these is the interaction between the mesoscale circulations produced by flow over mountains and the larger scales of motion. The traditional view of the orographic effect on large-scale circulation being understood in terms of vertical stretching or compression of vortex tubes yields the correct answer, roughly and qualitatively. Accurate forecasting, however, requires an accurate calculation of orographic effects, and this is not available. Hence, for example, forecast skill at and beyond 48 hours in the eastern United States is often limited by the quantitatively fallible predictions of the evolution of systems traversing the Rocky Mountains. In the plains states of America the problem is obviously even worse. Throughout the hemisphere, inadequacy of our physical understanding of large-scale orographic effects is a major obstacle to effective medium-range and extended-range forecasting.

The second topic might be called the mesoscale phenomenology of the surface boundary layer.

Much boundary-layer work has concentrated on the vertical fluxes of heat, water vapor, and momentum between surface and overlying air, with a view toward explaining the vertical structure of the layer and toward accounting for the fluxes between the boundary layer and the large-scale "free" atmosphere above. There seems to be an insufficient appreciation of the horizontal structures, often mesoscale, within the boundary layer itself; the coastal front is an example. In fact, we note that in this review "atmospheric boundary layers" and "mesoscale meteorology" are regarded as separate topics. We would hope that encouraging progress toward merger will have been effected by 1986, when the next Quadrennial Report is published.

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ATMOSPHERIC BOUNDARY LAYERS

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1. Introduction

Since more than a dozen papers relating to atmospheric boundary layers are published every month, it is necessary to restrict the scope of this review. Here we address only fundamental aspects of the physics of the boundary layer and exclude research related to instrumentation, specific applied areas of boundary layer research, and interactions with larger-scale circulations.

Some aspects of the boundary layer will be treated very briefly or included only in the reference lists while other topics, which have been omitted in past reviews, will be covered in more depth. Wyngaard (1978, 1979), Paegle (1979), Zeman (1981) and Nieuwstadt (1982) recently surveyed many aspects of atmospheric boundary layer research.

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2. Mixed-Layer Growth

The recent papers of Deardorff (1978), Artaz and André (1980), and Tennekes and Driedonks (1981) include surveys of mixed-layer growth models. Comparison between growth models with observations remains unsatisfactory because observational errors are probably greater than differences between model predictions. Uncertainties include estimation of advection, vertical motion and even mixed-layer depth. All three are often influenced by mesoscale and wave-like phenomena.

Mixed-layer growth models are sometimes calibrated with laboratory data. Annulus experiments show that turbulence may be partly suppressed by inertial stability. Then the entrainment occurs primarily in and near the side-wall boundary layers (Scranton and Lindberg, 1983; Deardorff and Willis, 1982). As a result,