Chapter 3

Advances in Knowledge and Understanding of Extratropical Cyclones during the Past Quarter Century: An Overview

Richard J. Reed

Department of Atmospheric Sciences, University of Washington, Seattle, Washington 98195

3.1 Introduction

The purpose of this chapter is to present an overview of the progress that has been made in knowledge and understanding of the extratropical cyclone in the roughly quarter-century that has elapsed since Palmén worked actively on the subject. It is recognized that other contributors to this volume will describe more fully Palmén's own contributions to the subject and will treat in greater detail various aspects of the subject that are only touched upon here.

With the purpose of keeping the overview to manageable size, it has been decided to focus on only certain aspects of the cyclone problem. Topics to be emphasized are the structures of fronts and cyclones and the processes of frontogenesis and cyclogenesis. Such important topics as the role of cyclones in the general circulation, orographic cyclogenesis and mesoscale precipitation features within cyclones will be left for others to discuss. With the purpose of putting the advances of the past quarter-century into perspective, the development of knowledge and understanding of the extratropical cyclone prior to 1960 will first be sketched.

3.2 Status of the Cyclone Problem Prior to 1960

As documented by Gisela Kutzbach (1979) in her treatise, *The Thermal Theory of Cyclones: A History of Meteorological Thought in the Nineteenth Century*, a considerable knowledge of cyclone structure and behavior existed prior to World War I and many relevant thermodynamic and dynamic principles were understood. Espy, Ferrel, Dove, Loomis, Buchan, Mohn, Ley, Köppen, Bigelow, Margules, von Ficker, Dines and Shaw are among the many early meteorologists whose substantial contributions are described in Kutzbach's book. The picture of cyclones gleaned from the efforts of these early investigators, however, seems fragmentary when viewed against the remarkable synthesis achieved by the Bergen school of meteorologists under V. and J. Bjerknes in the period following World War I. In the polar front theory of cyclones, which they put forth at that time (Bjerknes and Solberg 1922), the cyclone forms as a result of an instability of the polar front, a surface of discontinuity separating tropical and polar air masses. Beginning as a wave on the front, the cyclone undergoes a characteristic life cycle that terminates in the occluded stage in which the tropical air has been lifted aloft by the sinking and spreading of the polar air. The kinetic energy of the cyclone derives from the potential energy released in the rearrangement of the air masses.

These concepts are illustrated in Figs. 3.1 and 3.2, taken from the paper of Bjerknes and Solberg. The first figure depicts the structure of the developing cyclone, showing the arrangement and motion of the warm and cold air masses at the surface (middle panel) and the cloud and precipitation distribution in vertical sections taken to the north (upper panel) and south (lower panel) of the cyclone center. A broad region of cloud and precipitation is seen to form from the upgliding of the warm air along the sloping warm front. A narrower band is produced by the upthrusting of the warm air by the advancing cold front. The figure represents a slightly revised version of the diagram appearing in J. Bjerknes' (1919) pioneering paper on the subject. Figure 3.2 illustrates the life cycle of the cyclone from its birth as a wave on the polar front to its demise as a decaying vortex within the cold air.

Although a number of features of the polar-front theory remain essentially intact, many modifications and extensions have become necessary as new knowledge has been gained. Swarm or serial ascents by balloon-borne meteorographs, taken in the late 1920s and early 1930s, gave the first adequate picture of upper-air structure in cyclones. Some knowledge of the structure had been obtained earlier from cloud motions, from mountain observations, from occasional balloon and kite soundings and from manned ascents. The serial ascents revealed, not surprisingly, that the depiction of the polar front as a discontinuity surface separating tropical and polar air masses was an over-idealization. A narrow layer of transition constituted a more fitting (but still idealized) description of the thermal structure. The altered manner of depicting the polar front is illustrated in Fig. 3.3, taken from a familiar paper of Bjerknes and Palmén (1937).

The serial ascents also revealed the existence of an upper-tropospheric wave located over and rearward of the surface cyclone and the presence of considerable baroclinity in the warm air above the frontal surface. Thus Bjerknes (1937), in a theoretical paper published concurrently with the aforementioned observational study, recognized two components to cyclogenesis: the previously hypothesized frontal instability and the instability of the upper-level wave in the baroclinic westerlies. The underlying concepts were later elaborated in a joint paper with Holmboe (Bjerknes and Holmboe 1944).

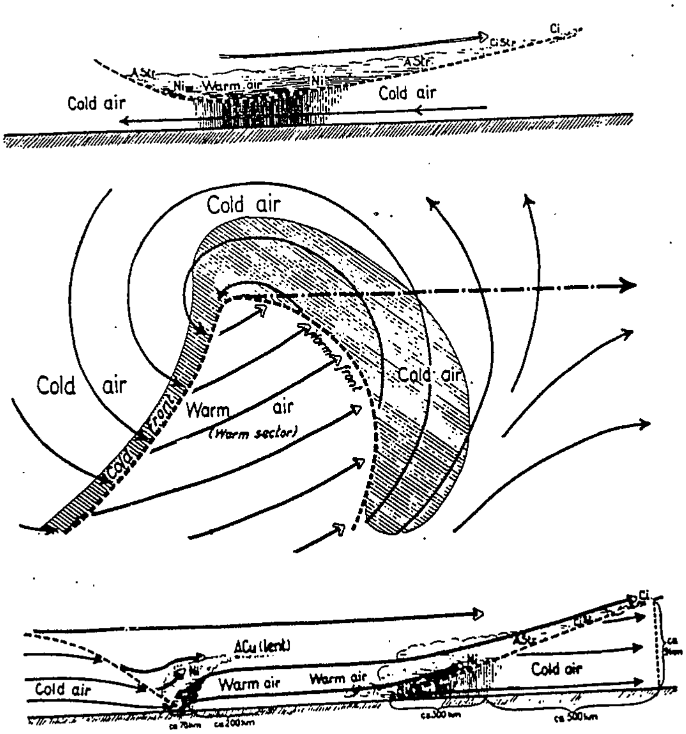

FIG. 3.1. Idealized cyclone (Bjerknes and Solberg 1922).

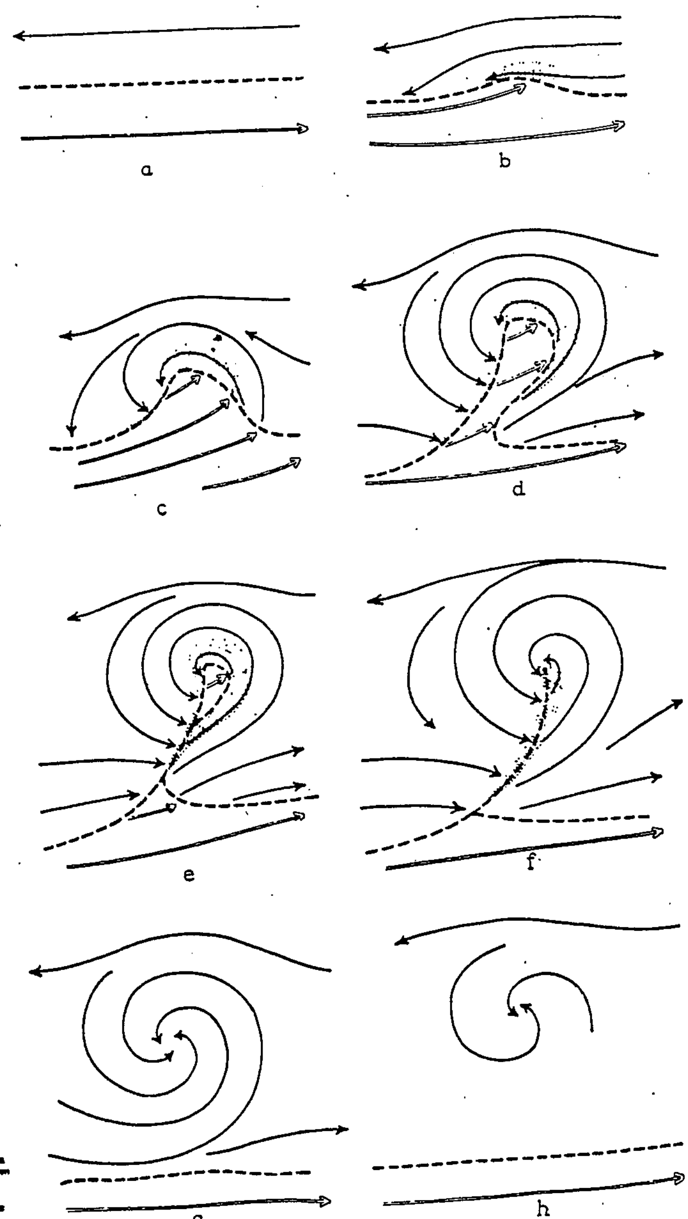

FIG. 3.2. The life cycle of cyclones (Bjerknes and Solberg 1922). See also Fig. 10.12.

With the establishment of radiosonde networks in the late 1930s and 1940s, the stage was set for even more rapid advance in defining the three-dimensional structure of the atmosphere. Particularly notable was the work of the Chicago school under the leadership of Rossby and Palmén (see Chapter 2). Many of the ideas and findings of this school were summarized in a famous paper which they coauthored with nine other distinguished meteorologists (Staff Members 1947). This paper introduced the concept of a jet stream and related the new entity to frontal and tropopause structure. In discussing the meanderings of the upper westerlies (and jet stream) the authors distinguished between slow-moving, long waves with lengths of 50 to 120 degrees of longitude and shorter frontal waves that moved through them and were steered by them. Defant (1912) had earlier speculated on the existence of long circumpolar waves based on his finding of a 5.7-day periodicity in rainfall records, and Bjerknes and Solberg related their idea of cyclone families to the hypothesized waves. The Chicago investigators also discussed the important phenomena of cutoff lows and highs (or blocks) and of downstream development caused by energy dispersion. A feature of the Chicago school, particulary evident in Palmén's work (1951), was the view of synoptic disturbances as vital elements in the exchange processes of the general circulation (Section 1.4). This view also had antecedents in the work of the Bergen school (e.g., Bjerknes and Solberg 1922).

The discovery of the jet stream tended to draw attention away from thermal discontinuities, whether in the temperature itself or in the temperature gradient, and to focus it instead on less sharply bounded zones of enhanced baroclinity. Such a zone is illustrated in Fig. 3.4 which shows a mean cross section through the jet stream on a particular day over North America (Palmén and Nagler 1948). It is evident that the heavy lines, representing the boundaries of the principal frontal layer, are not surfaces of discontinuity and that a substantial portion of the temperature contrast between the polar and tropical regions resides outside the layer.

Simultaneously with the expanded view of atmospheric structure came a theoretical development of first magnitude: the introduction of the theory of baroclinic instability by Charney (1947) and Eady (1949). This theory shifted emphasis away from frontal discontinuities as the seat of the instability responsible for cyclogenesis to the instability of the broad baroclinic westerlies. The theory successfully predicted the structure of the incipient waves, gave realistic growth rates for their development and accounted for their characteristic wavelengths. The success of the baroclinic theory, coupled with the failure of earlier works on frontal instability (Solberg 1928; Kotschin 1932; Bjerknes and Godske 1936) to obtain a satisfactory solution to the cyclone problem, resulted in general acceptance of baroclinic instability as the fundamental cause of cyclogenesis.

A further change in perspective regarding fronts occurred in the 1950s with Phillips' (1956) first successful numerical simulation of the atmospheric general circulation. In this experiment, started from a simple initial state, baroclinic disturbances formed which later acquired front-like features. Thus the typical frontal structure was seen as a consequence of the cyclogenesis rather than as a cause. Of course, this finding did not change the observational fact that in nature preexisting fronts are a favored seat of cyclogenesis. Cyclones can create fronts that in turn aid in the formation of new cyclones (Eliassen 1966). Actually, this idea was also contained in Bjerknes and Solberg's (1922) description of the cyclone family which required a "mother" cyclone to initiate the wave developments. Later numerical studies by Edelmann (1963), Hoskins and West (1979) and Takayabu (1986), begun from simple initial states featuring continuous but concentrated thermal fields and associated overlying jet streams, have confirmed Phillips' finding. Phillips' early

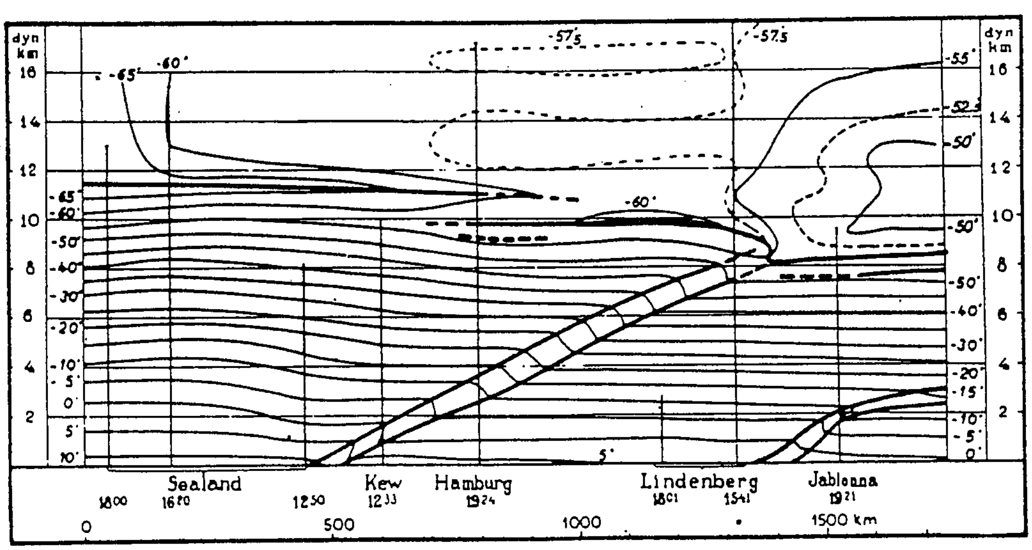

FIG. 3.3. "The 'zonal cross-section,' evening of the 15th of February, 1935" (Bjerknes and Palmén 1937).

picture, with fronts subjectively added, is reproduced in Fig. 3.5.

As the view of the polar front as a semipermanent discontinuity surface diminished and sharp fronts came increasingly to be regarded as transient features of the circulation, interest grew in the process of frontogenesis. Studies by Reed and Sanders (1953), Newton (1954), Reed (1955) and Sanders (1955) looked at the frontogenetical process in actual cases and in three dimensions. These studies made use of kinematical concepts developed by Bergeron (1928), Petterssen (1936) and Miller (1948) and also took account of vorticity considerations (Petterssen and Austin 1942). The studies of Reed and Sanders (1953) and Reed (1955) revealed that, in certain cases at least, strong upper-level frontal zones contained air with potential vorticity values typical of stratospheric air, suggesting an immediate stratospheric origin. The studies also revealed that in such cases a tilting of the isentropic surfaces by differential vertical motion, rather than the classical deformation process, was the prime mechanism in the frontogenesis. Sanders' work (1955) called attention to the extreme intensity that surface fronts can achieve, the rapidity with which the frontal intensity diminishes away from the surface, and the great strength of the frontogenetical process next to the ground, even under apparently steady-state conditions. It does no discredit to his work to state that part of its novelty stemmed from the excessive hold that the idealized picture of a more or less permanent and pervasive polar front had on the meteorological thinking of that era.

The idea that frontal zones are better regarded as regions of active frontogenesis than as semipermanent phenomena had another important consequence. It provided the impetus for Sawyer (1956) to formulate a diagnostic equation that can be used to measure the cross-front, ageostrophic circulation that forms in response to a prescribed frontogenetical forcing. As modified by Eliassen's elegant treatment (1962), this equation, which has come to be known as the Sawyer–Eliassen equation, has received wide application in later treatments of the frontal problem. The physical basis of the equation, first enunciated by Namias and Clapp (1949), is illustrated in Fig. 3.6.

This account of the pre-1960 period would not be complete without mention of the work of Sutcliffe (1947) and Petterssen (1956, pp. 320–339) on the problem of cyclone development. Their objective was to diagnose the factors conducive to cyclogenesis in the complex, finitely disturbed initial states that prevail in the real atmosphere, rather than to determine the growth of small perturbations in highly simplified flows, as done in theoretical studies. The provocative ideas of B type development (Petterssen et al. 1955) and A type development (Petterssen et al. 1962) have been an outcome of Petterssen's work. The former type refers to cyclone development at low levels associated with the approach of a preexisting upper-level trough to a low-level baroclinic zone. The latter type applies to development that takes place in a region of maximum baroclinity with the major contribution coming from the low-level thermal advection rather than from the vorticity advection aloft. Sutcliffe's expression for development, incidentally, was a forerunner of the much used quasi-geostrophic ω-equation (Fjørtoft 1955; Holton 1979, pp. 136–139; Trenberth 1978) which in Q-vector form (Hoskins et al. 1978) has received renewed interest.

3.3 Advances of the Past Quarter Century

As evident from the foregoing account, the post-1960 period began with a substantial knowledge of the major structural features of fronts and cyclones; with some

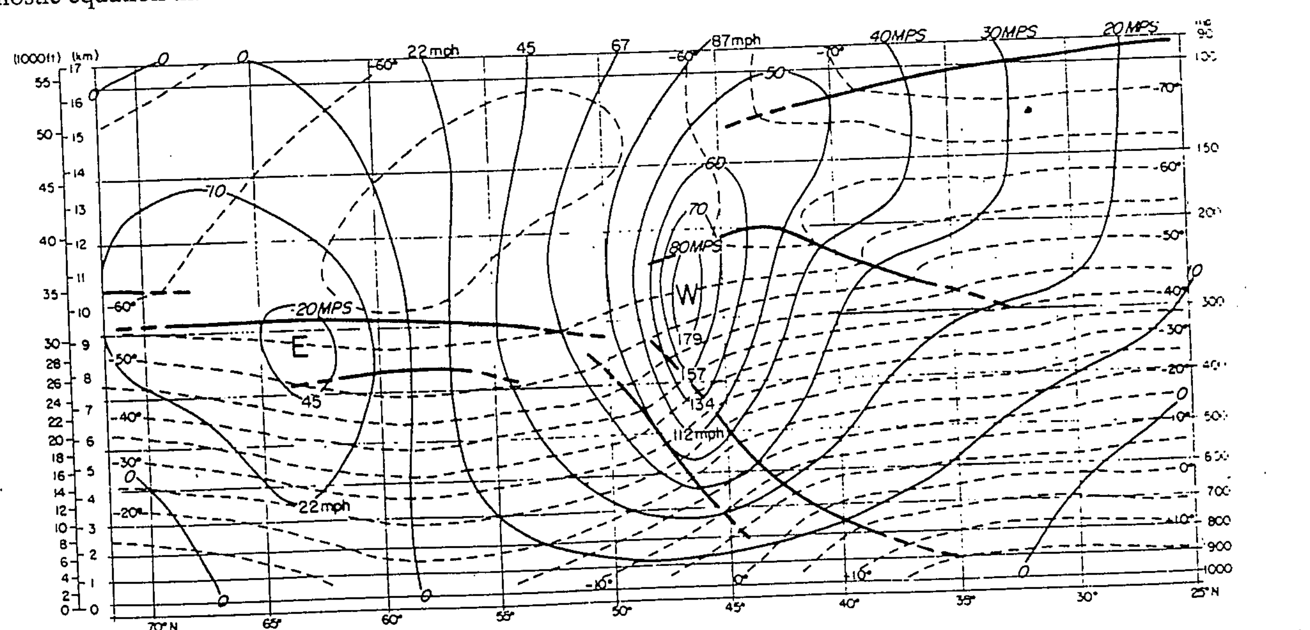

FIG. 3.4. A mean cross-section for 03 UTC 30 November 1946, showing the geostrophic westerly wind (solid) and temperature (dashed) over North America in a case of westerly flow. Heavy solid lines indicate boundaries of principal frontal layer and tropopause surfaces (Palmén and Nagler 1948).

understanding of frontogenetical processes and their role in frontal circulations; with an accepted instability theory of cyclogenesis (baroclinic instability); and with diagnostic tools (basically the quasi-geostrophic ω-equation) for the study of cyclone development in the real atmosphere. What have been the accomplishments of the past 25 years? From the standpoint of the present overview, the two most impressive advances have been in the areas of (1) theoretical understanding of frontogenesis and (2) numerical modeling of cyclones and cyclogenesis in real situations. These and other important advances of the past quarter-century will now be reviewed.

3.3.1 Surface Fronts and Frontogenesis

As summarized in a review article by Hoskins (1982), theoretical understanding of frontogenesis advanced rapidly in the late 1960s and early 1970s, thanks largely to the work of Stone (1966), Williams (1967, 1972), Hoskins (1971) and Hoskins and Bretherton (1972). In these works, the formation and evolution of fronts were studied on the basis of either a prescribed deformation field or a deformation field produced within an amplifying baroclinic wave. Realistic frontal structures featuring extreme thermal concentrations next to the surface, reminiscent of Sanders' (1955) observational results, were obtained in several of these studies. Stone's analytic solution, based on a prescribed deformation field and the quasi-geostrophic equations, succeeded in producing a sharp front at the surface but contained a number of unrealistic features, including the lack of a frontal slope, an unrealistic vorticity field, and a thermal gradient that grew exponentially with time so that an infinite time was required for the surface discontinuity to develop. Stone hypothesized that the lack of a frontal slope and the unrealistic vorticity distribution were consequences of the neglect of nonlinearities in the quasi-geostrophic solution, and further hypothesized that the effect of an induced cross-front, ageostrophic circulation would be to tilt the isentropes and thereby produce a realistic thermal wind field and associated vorticity distribution. This hypothesis was borne out by the later work of Williams (1967), in which the primitive equations were solved numerically, and in the work of Hoskins and Bretherton (1972), in which analytic solutions were obtained with use of the geostrophic momentum or semigeostrophic equations (see Section 9.8). A direct comparison of quasi-geostrophic and nongeostrophic frontogenesis was made by Williams (1972).

In addition, Williams' (1967) numerical results suggested that frontal discontinuities can form in finite time in the primitive equation system. Hoskins and Bretherton (1972) proved the correctness of this proposition in the case of the geostrophic momentum equations. Thus the key idea has emerged from these papers that frontogenesis may be regarded as a two-step process, in which the geostrophic deformation field of the growing baroclinic wave first concentrates the thermal gradient, and the induced secondary frontal circulation then causes a "collapse" of the thermal gradient to a true discontinuity near the ground.

Of course, when observed at sufficiently fine resolution, fronts in the real atmosphere always appear as transition zones of finite width, suggesting that some process acts to limit the intensity of the thermal gradient and thereby to bring about a steady-state condition. This process is generally considered to be turbulent diffusion (Welander 1963; Williams 1974). In a numerical experiment that included diffusion, Williams found that fronts can indeed form in a day or two and then persist in a quasi-steady state

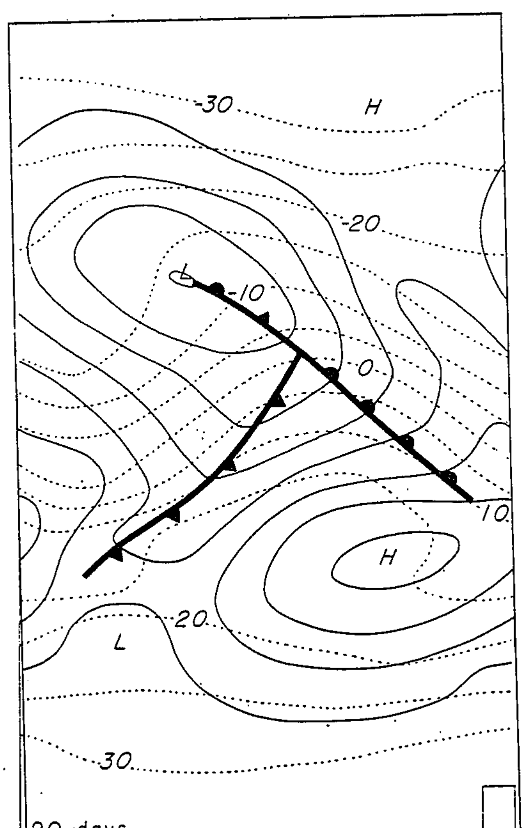

FIG. 3.5. Distribution of 1000 mb contour height at 200-ft intervals (solid lines) and 500-mb temperature at 5°C intervals (dashed lines) at 20 days (Phillips 1956). Fronts (added) have been subjectively located on the basis of the contour and isotherm configurations.

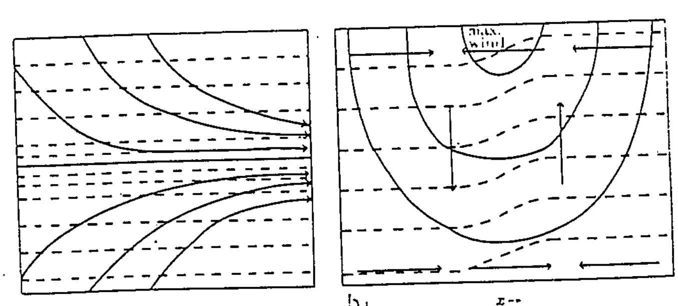

FIG. 3.6. (a) Horizontal distribution of streamlines (solid) and isotherms (dashed) in a frontogenetic confluence. (b) Vertical section showing confluent flow and the induced transverse ageostrophic circulation required to maintain thermal wind balance under the frontogenetical action (Sawyer 1956).

in which the frontogenetical processes are balanced by diffusive processes.

On the other hand, Orlanski et al. (1985) have argued that as the frontal scale becomes even moderately small, nongeostrophic accelerations (not contained in semigeostrophic theory) may lead to a reduction of the vorticity generation, thereby limiting the frontal strength. Gall et al. (1987) have recently tested this hypothesis in a nonhydrostatic, nondiffusive, primitive equation model in which the horizontal grid size was systematically reduced from 20 km to 280 m and the vertical resolution reduced from 320 m to 40 m. They found no evidence of a physical process that limits frontal collapse in the absence of diffusion, though their experiments did show some tendency for the line of maximum vorticity and line of maximum convergence to separate with time, as happened in the numerical experiments of Ross and Orlanski (1982) and Orlanski and Ross (1984).

Paradoxically, although turbulent diffusion appears to be the main mechanism limiting the frontal scale, surface friction, as demonstrated long ago by Eliassen (1959), strengthens the transverse frontal circulation near the ground and thereby contributes to surface frontogenesis. Eliassen's qualitative arguments have since been given quantitative support by aircraft measurements taken within a fully developed, quasi-steady front over the ocean (Bond and Fleagle 1985), though Eliassen (1966) later acknowledged some weaknesses in his earlier arguments (see Section 9.5). Bond and Fleagle evaluated indirectly the total forcing of the cross-front circulation by substituting observed data in the left-hand side of the Sawyer-Eliassen equation (see Section 9.2, Eqs. (9.12) and (9.13)). They then compared the inferred forcing with the forcing derived from the observed geostrophic wind and temperature fields, the forcing that ordinarily appears in the equation and estimates of the additional forcings caused by condensation heating and boundary-layer friction. The inferred forcing showed a strong maximum at 900 mb just ahead of the front (see vertical motion distribution in Fig. 8.23). A similar, but somewhat weaker, maximum appeared in the directly computed total forcing. This mainly resulted from the frictional component.

The effect of latent heat release in clouds on frontogenesis has been investigated in numerous studies (Hoskins and Bretherton 1972; Williams et al. 1981; Mak and Bannon 1984; Hsie et al. 1984; Ross and Orlanski 1978; Baldwin et al. 1984; Hsie and Anthes 1984; Bannon and Mak 1984). These studies have shown that condensation heating has a substantial enhancing effect. This result is not surprising in view of Sawyer's (1956) and Eliassen's (1959) early demonstrations that latent heat release strengthens the cross-front ageostrophic circulation.

It is evident from the Sawyer-Eliassen equation that small static stability or, more generally, small symmetric stability (Hoskins 1974) can also have a substantial enhancing effect on the secondary frontal circulation. The effect of small symmetric stability was first demonstrated by Hoskins and Bretherton (1972) in an example that assumed the potential vorticity to be everywhere zero. Such a condition corresponds to a symmetrically neutral state. The neutral state is rarely seen in a dry atmosphere but, as these authors and others (e.g., Kleinschmidt 1941; Sawyer 1949; Bjerknes 1951; Bennetts and Hoskins 1979; Emanuel 1983; Sanders and Bosart 1985) have pointed out, it can readily be achieved along moist isentropic surfaces in cloudy air. Studies by Emanuel (1985) and Thorpe and Emanuel (1985) show the great narrowing and intensification of the frontal upglide that occurs when the moist symmetric stability is small. Their results were anticipated in a little-known work of Todsen (1964), discussed and illustrated in Section 9.7.

On the observational side, the most notable advance in knowledge of frontal structure has been the documentation from aircraft, tower and radar wind-profiler data of the remarkable intensity that fronts can achieve in the vicinity of the ground. Sudden wind shifts and sharp temperature drops with passage of cold fronts have been recognized by meteorological observers for a century or more, and examples of these can readily be found on strip charts from continuously recording instruments. Nevertheless, recent measurements of Shapiro (1984), Shapiro et al. (1985) and Bond and Fleagle (1985) have added significantly to the picture of frontal structure in the boundary layer. In particular, their results show that intense cold fronts have the appearance of the density currents seen in laboratory experiments and in their atmospheric counterpart, the squall line. A striking feature is the narrow plume of rapidly ascending air (width on the order of a kilometer or less and speed on the order of 5–10 m s^{-1}) that occurs at the nose of the front (see Figs. 8.23 and 10.9). The plume has also been measured by Doppler radar (Browning and Harrold 1970; Carbone 1982; Hobbs and Persson 1982) and is responsible for what is termed the narrow cold front band in the Houze and Hobbs (1982) classification of mesoscale rainbands. Cloud and precipitation, however, are not essential for its formation, as evidenced by its occasional occurrence in dry air (Shapiro 1984). Diagrams from Shapiro's paper are shown in Fig. 3.7. The upper panel illustrates the extreme sharpness of the front. Only ten seconds were required for the front to pass the instrumented tower from which the measurements were taken. At the observed frontal velocity of 17 m s^{-1} this time interval corresponds to a frontal width of 170 m.

The lower panel shows the vertical velocity measured at the front. Already at 300 m the velocity exceeded 5 m s^{-1} in the warm air just ahead of the narrow transition zone. It is perhaps of significance that the warm air ahead of the front and the cold air to the rear were highly unstable, possessing superadiabatic lapse rates from the heated ground to the uppermost level. Keyser and Anthes (1982) obtained a similar but much weaker plume in a numerical experiment with horizontal grid resolution of 20 km that incorporated boundary layer physics. Their diagnosis

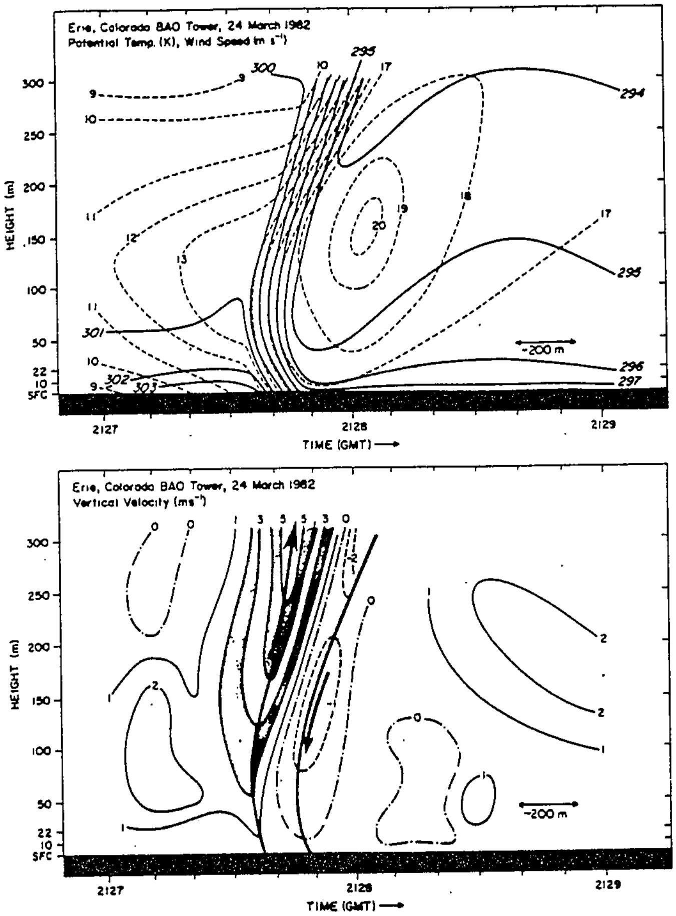

FIG. 3.7. Upper panel: Cross sectional (time series) analysis of BAO wind tower temperature and wind measurements. Lower panel: Corresponding analysis of vertical motion. Note the distance scale in lower right (Shapiro 1984).

revealed that the jet-like feature, as in the case study of Bond and Fleagle (1985) and in a later model simulation of Baldwin et al. (1984), was a consequence of surface friction.

Also worthy of mention on the observational side are the studies of the phenomenon of coastal frontogenesis by Bosart et al. (1972), Bosart (1975, 1981) and Keshishian and Bosart (1987). These highlight an additional mechanism that can produce substantial frontal circulations—the differential heating that takes place between land and sea when a cold polar anticyclone moves offshore. The process is aided by elevated terrain that traps the cold air in the vicinity of the coast and by latent heat release in cumuliform clouds that form over the warm waters. The most pronounced cases of coastal frontogenesis occur along the Carolina coast of the eastern U.S. where the coastal configuration, in combination with the proximity of the warm Gulf Stream waters and elevated terrain of the Appalachian Mountains, create an unusually favorable set of conditions. Some success in modeling the phenomenon has been achieved by Ballentine (1980), Stauffer and Warner (1987) and Uccellini et al. (1987).

3.3.2 Upper-Level Fronts and Frontogenesis

Prior to the early 1960s some doubts existed regarding the validity of the tropopause folding concept (Reed 1955), in which a tongue of stratospheric air was envisaged to descend into the upper or middle troposphere and assume the character of an upper-level frontal zone. These doubts were largely dispelled by aircraft observations made by Briggs and Roach (1963), who measured humidity and ozone mixing ratios near jet streams and found in some cases protrusions of stratospheric air with low water vapor and high ozone content within sharply-bounded frontal zones beneath the jet stream core. Briggs and Roach's results were reinforced a short time later when Danielsen (1964, 1968) succeeded in directing aircraft carrying radioactivity measuring equipment through an intense upper-tropospheric frontal zone. Having been deposited in the stratosphere in nuclear weapons tests, the radioactivity, like ozone, served as a tracer of stratospheric air.

Some of Danielsen's results are reproduced in Figs. 3.8 and 3.9. Figure 3.8 depicts traces of temperature and accumulated total β activity measured at two flight levels during back-and-forth transects through the frontal zone.

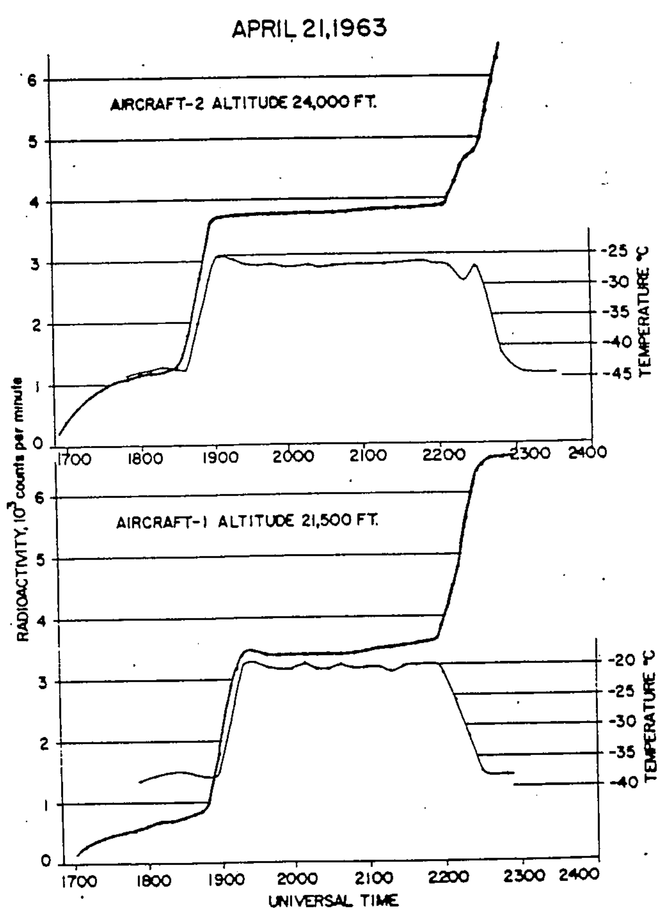

FIG. 3.8. Continuous traces of accumulated total β activity (cpm) and temperature measured along WB-50 flight paths in Fig. 3.9 (Danielsen 1968).

A nearly perfect match is seen between the two traces. On both warm and cold sides of the zone, the radioactivity is low (small slope of the curve) signifying tropospheric air. Within the zone, the extraordinary intensity of which is evidenced by the nearly 20°C temperature change across it, the radioactivity is vastly greater (steep slope). The presence of stratospheric air is proved beyond doubt.

A cross section through the tropopause fold, depicting β activity of strontium-90 and potential vorticity, appears in Fig. 3.9. Also shown are the flight paths of the WB-50 aircraft. The stratospheric extrusion is plainly evident in both the radioactivity and potential vorticity. Other equally remarkable tropopause folds have been documented in later penetrations of upper-level fronts by instrumented aircraft (Danielsen and Mohnen 1977; Shapiro 1978, 1980; Danielsen et al. 1987). Bosart (1970) has made an in-depth study of an extreme case that occurred at a time when a special series of three-hourly radiosonde ascents were taken in the eastern U.S. In the works cited, a variety of tracers have been employed including radioactivity, ozone, carbon monoxide, water vapor, condensation nuclei and potential vorticity.

Since, in general, isentropic potential vorticity is not conserved in the presence of diabatic heating and viscous and other nonconservative body forces (Staley 1960), it cannot be expected to be as conservative as some of the other tracers. In particular, Shapiro (1976, 1978) has presented examples of abnormally large stratospheric potential vorticity values, immediately on the cyclonic side of the jet core (see Fig. 10.4), that have no counterparts in the ozone distribution. He has explained these anomalous amounts in terms of the vertical divergence of turbulent heat fluxes above and below the jet core, but his interpretation has been questioned by Danielsen et al. (1987). Because of the issue of stratospheric-tropospheric exchange, the subject of turbulent mixing in the vicinity of tropopause folds has attracted much interest, but the topic is beyond the scope of this overview.

Diagnostic studies of upper-level frontogenesis have focused on the problem of explaining the cross-stream indirect thermal circulation that observational studies (e.g., Reed and Sanders 1953) have shown is essential to the process. Two mechanisms have been proposed: (1) the mechanism introduced by Palmén and Nagler (1949), and recently revived by Newton and Trevisan (1984a), in which an indirect circulation forms in a plane normal to the jet stream as air flows from an upstream ridge to a downstream trough under the constraint of gradient wind balance; and (2) the mechanism postulated by Shapiro (1981), based on the Sawyer–Eliassen equation, in which an along-front thermal gradient within an essentially straight frontal zone produces a shift of the downward branch of the transverse circulation from the cold to the warm side of the zone. The underlying mechanism was first discussed by Eliassen (1962) with reference to lower-level warm frontal circulations. In the upper-level case the along-front thermal gradient is such that cold advection occurs (Shapiro 1983). Uccellini et al. (1985) applied the Sawyer–Eliassen equation to an actual case of tropopause folding, successfully reproducing the measured transverse circulation while isolating the effect in question.

As described by Bosart (1970) and Shapiro (1983), most cases of intense upper-level frontogenesis occur when a short-wave trough moves in a northwesterly current from an upstream long-wave ridge to an amplifying (or "digging") downstream long-wave trough. Thus the two mechanisms (if they can be regarded as dynamically independent) reinforce each other (see Section 10.2.2). Gen-

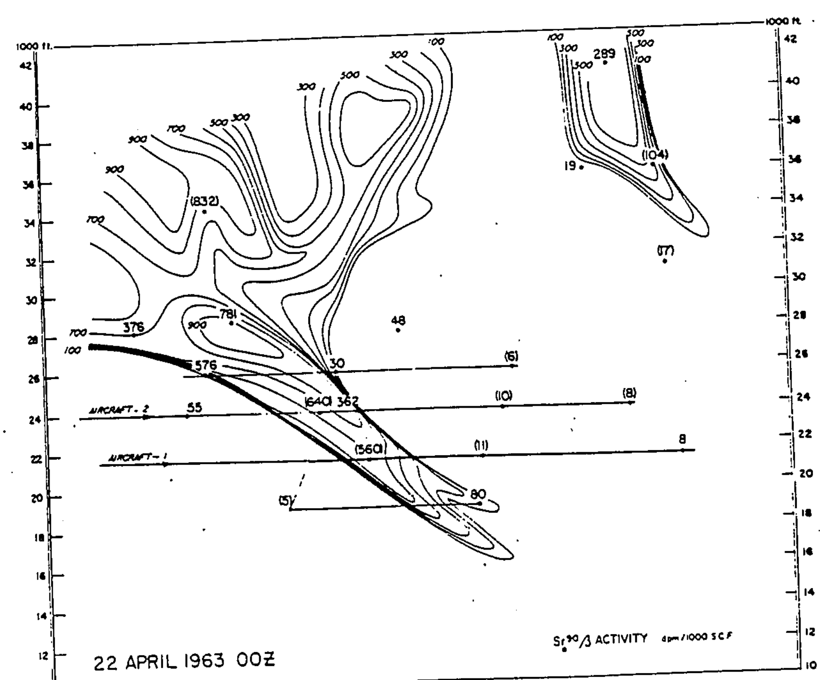

FIG. 3.9. Potential vorticity (contoured at intervals of 100×10^{-9} m s K kg^{-1}) and β activity of strontium-90 (dpm/KSCF) (Danielsen 1968).

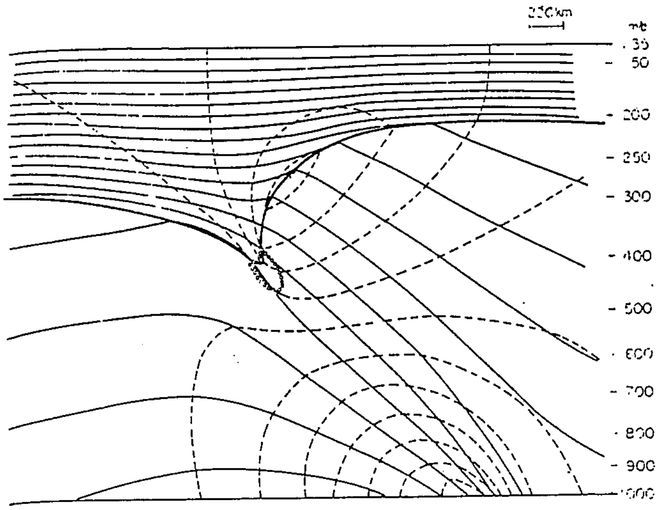

FIG. 3.10. Upper-level frontogenesis in a two-region semigeostrophic model. Solid lines, potential isotherms (interval 7.8 K); dashed lines, isotachs (interval 10.5 m s^{-1}). Richardson number contour value of 1.0 is indicated by small circles (Hoskins 1971).

erally the strongest thermal gradient is achieved in the vicinity of the trough axis, but occasionally the zone of maximum intensity advances somewhat ahead of it. Examples may be found in Reed (1955) and Bosart (1970).

Limited success has been achieved in simulating upper-level fronts and tropopause folds with two-dimensional dynamical models. The solution to the analytic model of Hoskins (1971) and Hoskins and Bretherton (1972), in which a uniform horizontal deformation field was allowed to act on an initial temperature field, is shown in Fig. 3.10. The model uses the geostrophic momentum and Boussinesq approximations and contains a tropopause separating a tropospheric region of small uniform potential vorticity below from a stratospheric region of large potential vorticity above. The solution shows a rudimentary tropopause fold. Another noteworthy feature is the distinct separation of the upper- and lower-level fronts. A subsequent nonBoussinesq solution (Hoskins 1972) yielded an even more pronounced penetration of the stratospheric air into the troposphere, but neither simulation gave a fold of the magnitude seen in real cases.

Buzzi et al. (1981), using a two-dimensional semi-geostrophic model in isentropic coordinates, extended Hoskins' work to include a continuous potential vorticity distribution. A jet structure was prescribed, and upper and lower boundaries were chosen to be isentropic. An internal upper-tropospheric front formed under these conditions in response to the imposed deformation field but, as the authors pointed out, the circulation transverse to the front was direct, not indirect as found in observational studies. Moore (1987) has also studied the effect of a continuously-varying potential vorticity. According to Shapiro's (1981) diagnosis, however, an indirect frontal circulation can occur in the two-dimensional case provided there exists an along-front thermal gradient with cold advection within the zone. This idea has been tested by Keyser and Pecnick (1985) in a two-dimensional primitive equation model patterned after the work of Hoskins and Bretherton (1972). Their results, shown in Fig. 3.11, confirm Shapiro's reasoning. The "cold advection" case (Fig. 3.11b) exhibits an upper-level frontal zone, after 24 hours of integration, that is stronger than that of the "pure confluence" case (Fig. 3.11a) and that has the maximum subsidence displaced toward the warm side of the zone.

A number of studies of upper-level frontogenesis have been performed using three-dimensional models (Mudrick 1974; Shapiro 1975; Buzzi et al. 1977; Hoskins and West 1979; Heckley and Hoskins 1982; Newton and Trevisan 1984b). In general these have produced realistic fronts, but none has succeeded in duplicating some of the extreme

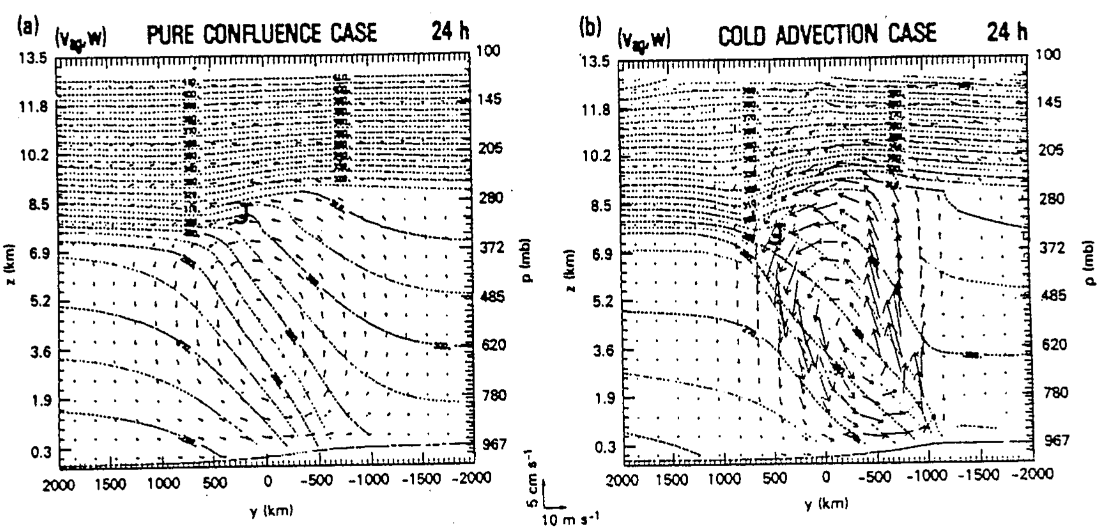

FIG. 3.11. Cross sections of the transverse ageostrophic circulation and potential temperature after the 24-h integration of a two-dimensional primitive equation model of frontogenesis due to (a) pure confluence and (b) confluence in the presence of upper-level cold advection (Keyser and Pecnick 1985). See also Fig. 10.7.

cases observed. It is possible that the oversimplified initial states used in the model simulations lack essential features of the initial states that are present in actual situations. For a more comprehensive discussion of upper-level fronts, the reader is referred to a recent review by Keyser and Shapiro (1986). A review of observational aspects of fronts, both upper and lower, can be found in Keyser (1986). Overviews of the subject are presented by Eliassen in Chapter 9 of this volume from the theoretical viewpoint, and by Shapiro and Keyser in Chapter 10 with emphasis on recent observational studies.

3.3.3 *The Cyclone as Viewed from Space*

Slightly more than a quarter-century has elapsed since a new observational tool of immense importance—the meteorological satellite—was introduced into meteorology. Cloud patterns seen in visible and IR images by the early polar-orbiting satellites have given a fresh perspective on cyclone structure and behavior, and the ability of satellites to view the entire globe has allowed cloud systems in cyclones over the oceans to be adequately documented for the first time. Geostationary satellites introduced later have provided time-lapse pictures that have proved particularly valuable in studying the evolution of cloud features and, with the help of water vapor imagery, moisture patterns as well (Petersen et al. 1984). Passive microwave radiometers have already shown promise in identifying integrated water vapor, cloud-water and rainwater patterns in cyclones, and doubtless will be used increasingly in both operations and research (McMurdie and Katsaros 1985).

What has been learned from this new observational tool regarding cyclone structure and evolution? One way to address this query is by posing a further question: How well has the classical picture of cloud organization put forth by members of the Bergen school (e.g., Bjerknes and Solberg 1921, 1922) stood up under the new stream of observations? Most investigators would agree that it has stood up very well, particularly over the oceans where the moisture supply is always sufficient to assure identifiable cloud masses. Numerous examples can be found in which the cloud patterns conform at each stage of development to the classical scheme. Cold, warm and occluded fronts are observed to have the expected signatures (WMO 1973).

On the other hand, numerous cases also exist in which the cloud patterns and their evolutions are complex and not easily fitted into the classical mold. In view of the expanded knowledge of cyclone structure that has been acquired since the classical picture was first propounded, the complexity is not surprising. The original cyclone model was founded on the idea of the cyclone as a wave on the polar front. Later came the idea of the upper baroclinic wave, and still later the concept of the jet stream. These new entities were viewed as being related to frontal waves and frontal zones, yet as having separate identities. Obviously the added circulation patterns associated with the upper waves and jet stream should be expected to complicate the simple picture derived from frontal considerations alone. Also the recognition that frontogenesis and cyclogenesis often occur in tandem rather than sequentially, and that interactions between more or less independent upper- and lower-level components are involved in many instances of cyclogenesis (Petterssen Type B), opens the possibility for a still wider range of structures. Clearly, modifications and extensions of the simple classical model are warranted and indeed have flourished (perhaps excessively) in the satellite literature.

An example of a modified scheme appears in Fig. 3.12. This illustrates the separation often made by satellite meteorologists between cirrostratus decks that lie on the warm or anticyclonic-shear side of the jet axis, and frontal clouds, whose tops are located at discretely lower levels. Although it might be speculated that the presence of distinct upper and middle-to-lower cloud patterns is indicative of distinct components to the vertical motion field, it is possible that the initial moisture distribution also plays a crucial role in shaping the cloud patterns (Durran and Weber 1988). As yet no satisfactory explanation exists concerning why the cloud tops vary more or less continuously with height in some cases, as postulated in the Norwegian model, and why in other cases they are arranged in steps.

Another pattern of development, often seen in satellite pictures but not included in presatellite thinking, involves cyclogenesis in cold air masses rearward of the polar front and poleward of the jet stream axis. Occurring mainly over the oceans during the winter half-year, the disturbances in question are first seen in satellite imagery as regions of enhanced cumulus convection. Subsequently the convective elements spread and merge, and a more or less solid comma-shaped cloud mass emerges. The final shape is not unlike that of many developing frontal cyclones, but typically the cold air systems are smaller in dimension. A satellite picture of such a system is shown in Fig. 3.13.

As illustrated schematically in Fig. 3.14, the disturbances are known to form in the region of positive vorticity advection ahead of a secondary upper-level vorticity maximum (Anderson et al. 1969; WMO 1973) or, from an alternative viewpoint, in the left exit region of a jet streak. It has further been found (Reed 1979; Mullen 1979; Reed and Blier 1986a,b) that the region of formation is characterized by weak to moderate baroclinity, appreciable surface fluxes of heat and moisture, and a conditionally unstable lapse rate through a substantial depth of the atmosphere. The systems are believed to be closely related to the polar lows that have been much discussed by British and Norwegian meteorologists, and indeed they are often regarded as one category of polar low (Businger and Reed 1989).

When a cold air comma approaches sufficiently close to the polar front, it may induce a wave development on the latter (Anderson et al. 1969). In many instances the wave

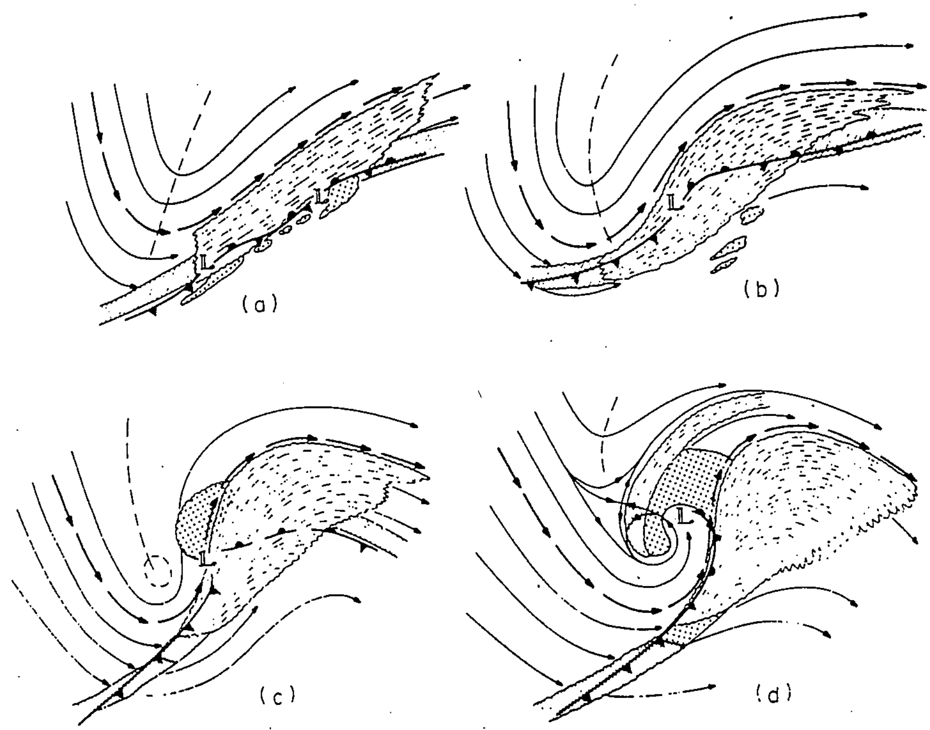

FIG. 3.12. Schematic diagram of the cloud patterns associated with an extratropical cyclone at four stages of development. Dashed lines indicate cirrostratus cloud decks, cross-hatching frontal cloud bands, and dots show low- or middle-level cloud decks. (Adapted from lecture notes of Roger B. Weldon by Wallace and Hobbs 1977, p. 262.) See also Fig. 8.11.

grows into a typical frontal cyclone, and the comma is rapidly absorbed into the frontal low. In other cases the wave and comma cloud maintain separate identities. Cases can also be found in which the polar low becomes the dominant partner (Reed 1979). In certain cases the merging of the two components have been described as "instant occlusion," the idea being that the comma supplies the low center on the poleward side of the jet axis and the frontal wave supplies the warm and cold fronts (Anderson et al. 1969). The process has been clarified in recent work by Browning and Hill (1985) and McGinnigle et al. (1988; described in Section 8.5). A case in which the merger of a comma cloud and frontal wave produced a major oceanic cyclone has been documented by Mullen (1983). The dual nature of the development is plainly evident in Figs. 3.15 and 3.16 which show satellite pictures and surface maps, respectively, of the evolving systems.

A method that has been found useful for understanding cloud patterns in cyclones is that of analyzing the flow on isentropic surfaces relative to a coordinate system moving with the storm (Kleinschmidt 1957, p. 135; Green et al. 1966; Browning and Harrold 1969; Carlson 1980). Browning presents a comprehensive discussion in Chapter 8. Provided the patterns are not undergoing rapid changes of shape, the streamlines of the relative flow represent air trajectories relative to the storm. Unfortunately, many of the most interesting situations involve rapid changes of shape, thereby complicating the interpretation. Furthermore, upper-air data needed for constructing such charts are available only at 12-hour intervals. A possible way around these difficulties is through use of trajectories constructed from the output of numerical prediction models that provide essentially continuous fields. Of course, to yield reliable information, the models must simulate accurately the development of the weather system under study, particularly the development of the cloud fields. Fortunately, the most advanced of the current operational and research models have reached the stage where meaningful diagnosis is possible.

3.3.4 Cyclogenesis

Theoretical studies of baroclinic instability have expanded greatly in number and scope during the past quarter-century. The pioneering studies of Charney and Eady employed linearized equations, a dry atmosphere and a basic flow characterized by vertical shear only. Later studies have utilized zonal jets with meridional as well as vertical shear (Pedlosky 1964a,b; Brown 1969a,b; Song 1971; Gall 1976a,b; Simmons and Hoskins 1976), have extended the analysis into the nonlinear regime (Simons 1972; Simmons and Hoskins 1978) and have treated moist processes (Gall 1976c; Mak 1982; Emanuel et al. 1987). It has been found that the inclusion of lateral shear, and hence of barotropic energy conversion processes, has only minor effects on the properties of the unstable baroclinic modes (e.g., Simmons and Hoskins 1976). The largest impact of the studies with zonal jets, particularly those extending into the nonlinear regime, has been in the knowledge gained regarding the interactions between the disturbances and the basic flow (Brown 1969b; Song 1971; Gall 1976b) and the variation of the interactions within the life cycle of the baroclinic wave (Simons 1972; Simmons and Hoskins 1978). The extended integrations with zonal jets have, as mentioned previously, yielded realistic cyclone and frontal evolutions (Edelmann 1963; Hoskins and West 1979; Takayabu 1986).

Condensation heating has been found by Gall (1976c) to enhance the growth rates of baroclinic disturbances at all wavelengths, but not to change the wavelength of maximum growth. Mak (1982) also found that condensation heating significantly increases growth rates but, contrary to Gall, obtained a significant decrease in the size of the most unstable wave. Of particular interest in connection with the effects of moist processes is the recent paper by Emanuel et al. (1987), in which they examine baroclinic instability in conditions of small stability to slantwise moist convection. Their results show the greatly enhanced growth rates and shortened wavelengths that are obtained as the moist potential vorticity, the measure of slantwise or symmetric stability, approaches zero. As remarked by the authors, in the semigeostrophic system of equations the potential vorticity acts as an effective static stability. It is the equivalent of the static stability in the quasi-geostrophic system. In this connection, previous studies (Staley and Gall 1977; Duncan 1977) using dry quasi-geostrophic models have revealed the scale-shortening effect of reduced static stability in the lower troposphere. Hayashi and Golder (1981), in a general circulation experiment comparing dry and moist atmospheres, found that baroclinic disturbances were much more energetic in the moist atmosphere and attributed this result to the lesser static stability that prevails under moist conditions.

A preponderance of the studies cited above have dealt with perturbations of normal mode form. Farrell (1982, 1984), however, has pointed out that in order to represent the initial conditions that exist in the real atmosphere prior to cyclogenesis, for example the conditions associated with Petterssen Type B development, it is necessary to augment the discrete normal modes by a continuous spectrum of neutral modes. Transient growth depends on the continuous spectrum; only at later stages do the normal modes predominate. In later work Farrell (1985, 1989) has further explored his contention that an initial value approach, rather than a normal mode approach, is required to study cyclogenesis in the real atmosphere. Another aspect of nonmodal baroclinic wave growth has been studied by Simmons and Hoskins (1979) with the use of a nonlinear primitive-equation model. They demonstrate how a localized disturbance introduced in a baroclinically unstable flow produces downstream (and upstream) wave development through energy dispersion.

It was mentioned earlier that prior to the advent of baroclinic instability theory a number of attempts were made, but with only limited success, to solve the problem of the stability of frontal waves, that is, waves on an ideal surface of discontinuity. This problem was addressed later by Orlanski (1968) for a wide range of Rossby and Richardson numbers. Based on his analysis, Orlanski identified four distinct regions of instability that he associated with the names of Rayleigh, Helmholtz, Eady and Bjerknes. Mechoso and Sinton (1983) have shown that the Bjerknes mode results from a juxtaposition of Rayleigh and Eady modes, and Sinton and Mechoso (1984) have studied the nonlinear evolution of the frontal waves, using a two-layer, shallow-water model. They show examples of rather realistic looking frontal cyclones for the Eady and Bjerknes modes. The relevance of their results to the less than ideal frontal cyclones that occur in nature remains to be clarified. In this connection it should be mentioned that Moore and Peltier (1987), in examining the stability of realistic frontal zone/jet stream structures, found the existence of a short-wave branch of unstable normal modes that is distinct from the classical Charney-Eady modes. The fastest growing mode in the newly discovered branch has a wavelength of about 1000 km. It derives its energy from the baroclinic conversion process.

A novel approach to understanding cyclogenesis, based on the conservation of isentropic or Rossby-Ertel potential vorticity, has been put forth by Hoskins, McIntyre and Robertson (1985). Underlying their approach is the invertibility principle, earlier recognized and utilized by Kleinschmidt (1957, pp. 113–116), which states that, subject to specification of a balance condition and a suitable reference state, the potential vorticity distribution is sufficient to infer the fields of the other meteorological variables such as winds, temperatures and geopotential heights. The approach has much pedagogical appeal, in that for dry-adiabatic motion it links earlier and later atmospheric states through a single simple conservation principle. During moist-adiabatic motion the (dry) isentropic potential vorticity is not conserved, and application of the principle becomes more complicated.

The final noteworthy advance in knowledge and understanding of extratropical cyclones, and in many ways the most remarkable, is in the realm of numerical prediction of cyclone development. It is easy to lose sight of the fact that every day at forecast centers around the world our knowledge of cyclones (or, more precisely, of the laws that govern their behavior) is being tested in real situations. The general success of the forecasts in data-rich areas bears witness to the fact that the cyclogenetical process is indeed now well understood—at least by the computer! Sensitivity experiments, in which various physical processes are withheld in turn from a full-physics control forecast, offer a means for better human understanding of the processes involved (see Section 12.3.2). Such experiments are becoming increasingly common.

Because of the special importance of rapidly intensifying storms and the difficulty, at least until recently, of predicting them (Sanders 1987), explosive cyclogenesis—defined roughly as deepening in excess of 24 mb in 24 hours (Sanders and Gyakum 1980)—has attracted much attention in recent years. Sanders and Gyakum showed that

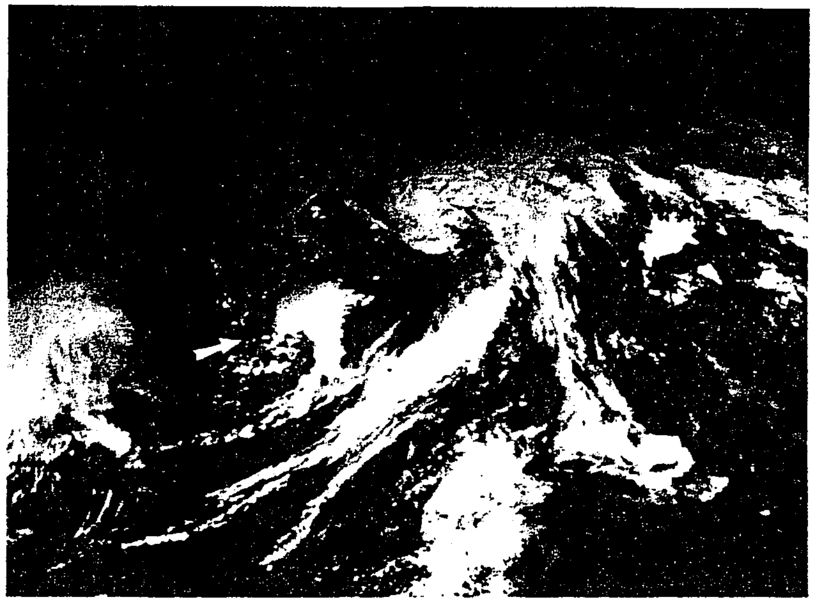

FIG. 3.13. Example of comma-shaped cloud pattern (indicated by arrow) in a polar air mass (Reed 1979).

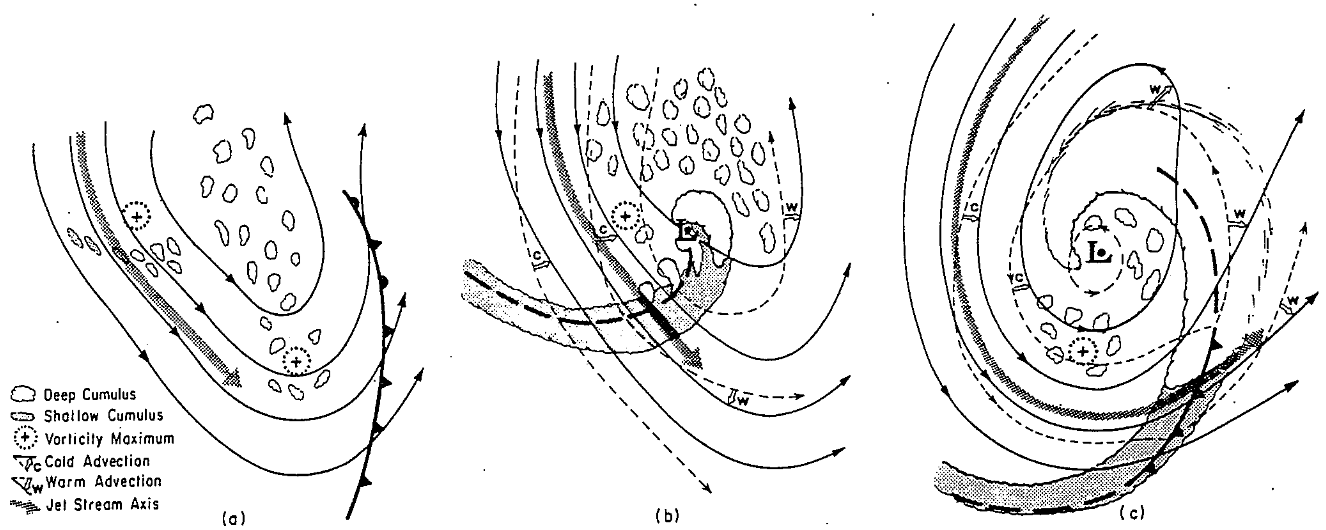

FIG. 3.14. Schematic diagram of comma cloud development: (a) incipient stage, (b) intensifying stage, (c) mature stage. Solid lines are 500 mb contours; dashed lines are surface isobars. Heavy broken line marks the surface trough. Frontal symbol in (c) indicates that the trough may assume frontal characteristics in some cases. Further explanation at lower left (Reed and Blier 1986a).

rapidly deepening storms are primarily maritime, cold season events and that they occur usually somewhat downstream of a mobile 500 mb trough within or poleward of the maximum westerlies and within or ahead of planetary-scale troughs. They found, furthermore, that such storms occur most frequently at the western ends of the Atlantic and Pacific storm tracks, preferentially near the regions of strongest SST gradients. They concluded from their study that explosive cyclones are baroclinic events aided by diabatic heating of an unidentified character. The baroclinic nature of explosive cyclones and their association with conditions that favor deep convection has been verified in a compositing study, based on weather ship observations, by Rogers and Bosart (1986).

Detailed case studies have been made of a number of extraordinary storm developments. Most studied has been the poorly forecasted Presidents' Day storm of 1979 which deposited a record-breaking snowfall on the middle Atlantic states. Bosart (1981) documented the crucial role of surface fluxes in moistening, heating and destabilizing the boundary layer and thereby establishing the coastal front upon which the low formed. He also documented the importance of an upper-level short-wave trough that advanced from the west and interacted with the coastal

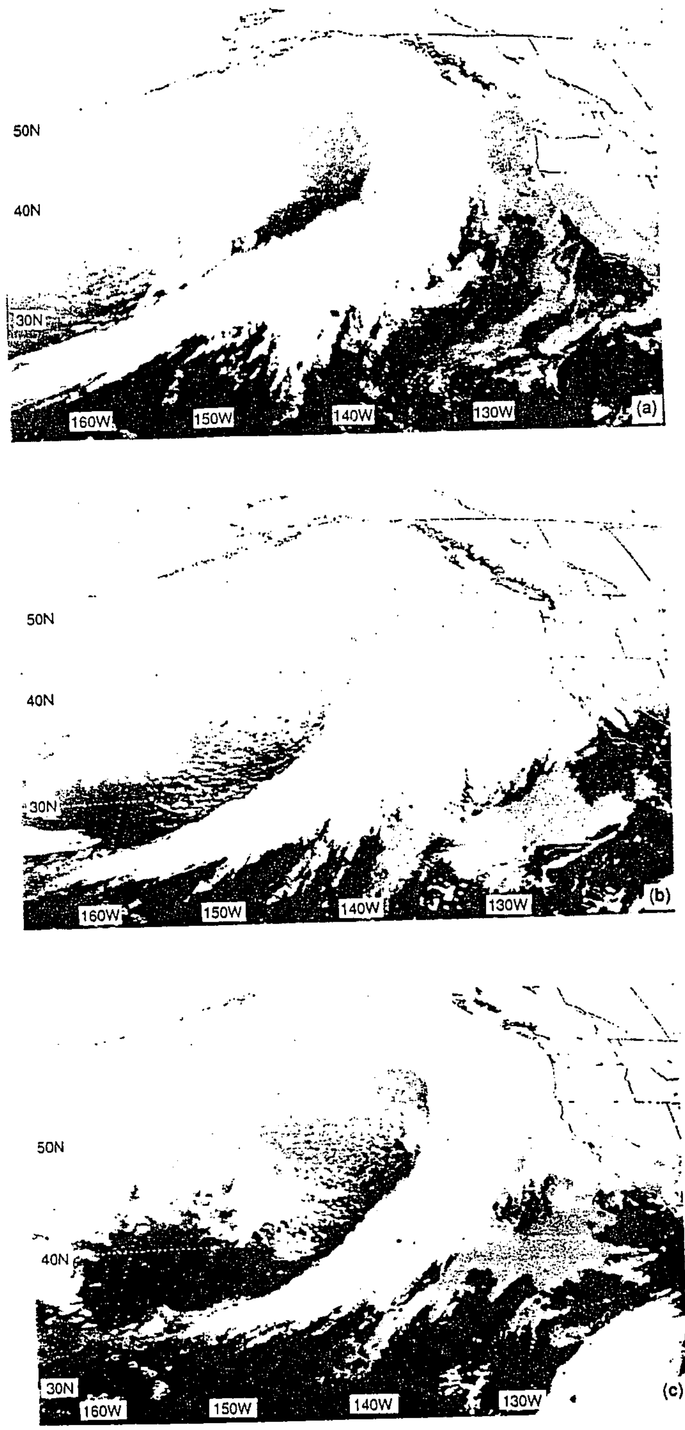

FIG. 3.15. GOES-W infrared (IR) satellite imagery at (a) 0545 UTC 7 January (b) 1745 UTC 7 January, and (c) 0546 UTC 8 January 1981 (Mullen 1983).

low, and the existence of convective activity in the vicinity of the storm center. A later diagnostic study of Bosart and Lin (1984) confirmed and extended the earlier results. Studies by Uccellini et al. (1984, 1985) called attention to the parts played in the development by an unbalanced subtropical jet streak, an induced ageostrophic low-level jet and the descent of stratospheric air in the tropopause fold beneath a polar jet streak that brought high potential vorticity air to mid-tropospheric levels upstream of, and prior to, the rapid deepening phase (see Section 6.3.2).

A second poorly predicted storm that has received extensive study is the so-called Queen Elizabeth II storm of September 1978 which exhibited a central pressure fall of more than 50 mb in 24 hours. Gyakum's (1983a,b) analysis of the storm led him to conclude that it originated as a shallow baroclinic disturbance and that its extreme development was a consequence of the release of latent heat in organized cumulus convection. He cited evidence of hurricane-like features in the storm. Uccellini (1986) questioned Gyakum's characterization of the QE II storm as a shallow baroclinic system, presenting convincing evidence that strong upper-level baroclinic features, including a pronounced tropopause fold, existed upstream of the storm prior to its intensification.

A storm in the eastern Pacific that deepened nearly 40 mb in 12 hours has been investigated by Reed and Albright (1986). The storm was an offshoot of an earlier low pressure system in the western Pacific that weakened as it passed through a long-wave ridge and then rejuvenated suddenly upon entering the succeeding downstream long-wave trough. It too was missed by the operational forecasts. Factors found to be associated with its development were strong, deep baroclinity, small static stability and possible symmetric instability within the frontal cloud band. Preconditioning of the environment by sea-surface fluxes immediately prior to the rapid deepening was also found to be a crucial factor in the development. Although a preexisting upper-level short-wave trough or jet streak was present upstream of the incipient surface low, the intensification of the upper-level vorticity pattern appeared to occur simultaneously with the storm development rather than prior to it.

Sensitivity studies, employing fine-mesh mesoscale models, have been conducted for each of these storms. Simulations performed by Uccellini et al. (1987) and by Atlas (1987) confirm dramatically the vital role of sea-surface fluxes in producing the coastal cyclogenesis that preceded the rapidly deepening phase of the Presidents' Day storm (see Secs. 6.3.3, 12.3.2g). Kuo, Shapiro and Donall (1990), in a moderately successful simulation of the QE II storm (see Fig. 10.14), initiated at the beginning of the 24-hour period of most rapid deepening, found a much lesser, but not inconsequential, role for the surface fluxes. Latent heat release was by far the major factor in the rapid deepening. Kuo and Reed (1988), in a simulation of the eastern Pacific storm, also initialized at the beginning of the rapid deepening stage, identified latent heat release as the dominant diabatic effect. During this stage surface fluxes had little impact on the forecast despite their importance in supplying heat and moisture at an earlier stage. The simulation also showed that a symmetrically neutral or slightly unstable state existed in the region of low-level inflow ahead of the storm where rapid spin-up of vorticity occurred. A narrow sheet of rapidly rising air (with velocities exceeding 50 cm s^{-1}) was located within

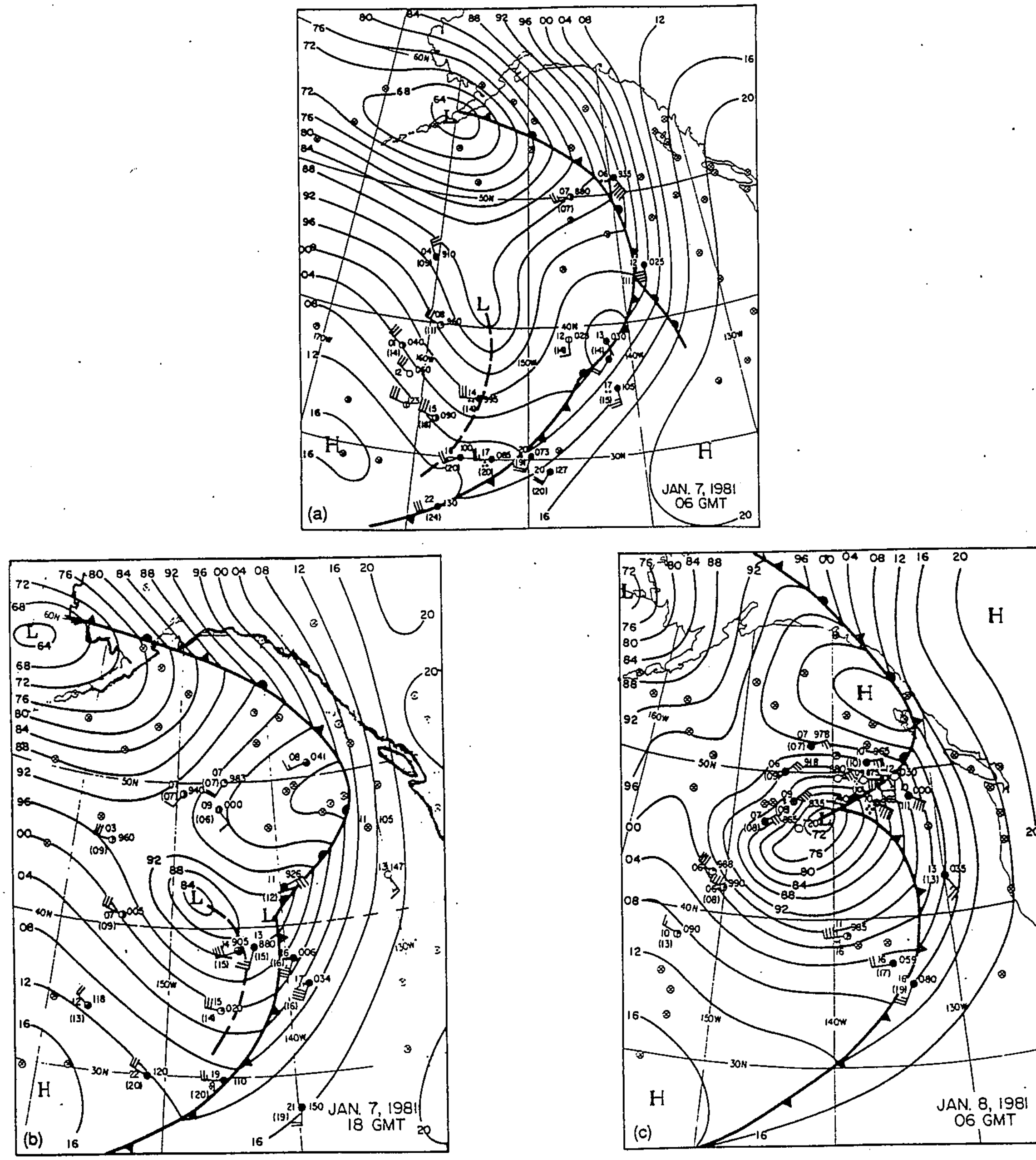

FIG. 3.16. Surface analysis for (a) 06 UTC 7 January (b) 18 UTC 7 January, and (c) 06 UTC 8 January 1981 (Mullen 1983).

the region. Their results suggest that a key element in rapid cyclogenesis is the strength of the slantwise ascent in and above the warm frontal (or warm frontogenetical) zone.

Danard and Ellenton (1980), Chen and Dell'Osso (1987) and Reed et al. (1988) have also performed illuminating sensitivity studies of rapidly deepening cyclones. These and other studies that are in progress should lead to a rapid increase in the understanding of processes that contribute to cyclogenesis in general and explosive cyclogenesis in particular.

3.4 Concluding Remarks

From the foregoing account it is apparent that the past quarter-century has witnessed a remarkable advance in knowledge and understanding of the extratropical cyclone, an advance made possible by the solid foundations laid down by earlier workers in the field such as the man we honor at this symposium—Erik Palmén. Thanks in part to the new opportunities created by the high-speed computer and in part to the ingenuity of a large number of skilled workers, much progress has been made in understanding how fronts and cyclones form and in predicting their development. On the observational side, new instruments and new and improved observing platforms—satellites, radars, research aircraft and towers—have allowed the cyclone both to be viewed in broad perspective and to be probed in fine detail.

Looking to the future, we can be optimistic. The advanced numerical models now in existence, and many of the new instrument systems, are yet to be fully exploited. Theoreticians still have new ideas to explore. When meteorologists gather a quarter-century from now to take stock of the progress that has been made on the cyclone problem, they will no doubt have a further proud record of accomplishment to look back on.

Acknowledgments

I wish to thank Brian J. Hoskins, Chester W. Newton and Louis W. Uccellini for reading the manuscript and offering many helpful suggestions and the National Science Foundation for supporting its preparation under grant ATM8421396A04.

REFERENCES

Anderson, R. K., J. P. Ashman, F. Bittner, G. R. Farr, E. W. Ferguson, V. J. Oliver and A. H. Smith, 1969: Application of meteorological satellite data in analysis and forecasting. *ESSA Tech. Rep. NESC51,* Government Printing Office, Washington D.C. [NTIS AD-697033].

Atlas, R., 1987: The role of ocean fluxes and initial data in the numerical prediction of an intense coastal storm. *Dyn. Atmos. Oceans,* **10**, 359–388.

Baldwin, D., E.-Y. Hsie and R. A. Anthes, 1984: Diagnostic studies of a two-dimensional simulation of frontogenesis in a moist atmosphere. *J. Atmos. Sci.,* **41**, 2686–2700.

Ballentine, R. J., 1980: A numerical investigation of New England coastal frontogenesis. *Mon. Wea. Rev.,* **108**, 1479–1497.

Bannon, P. R., and M. Mak, 1984: Diabatic quasi-geostrophic surface frontogenesis. *J. Atmos. Sci.,* **41**, 2189–2201.

Bennetts, D. A., and B. J. Hoskins, 1979: Conditional symmetric instability — a possible explanation for frontal rainbands. *Quart. J. Roy. Meteor. Soc.,* **105**, 945–962.

Bergeron, T., 1928: Über die dreidimensionale verknüpfende Wetteranalyse. *Geofys. Publ.,* **5**, No. 6, 111 pp.

Bjerknes, J., 1919: On the structure of moving cyclones. *Geofys. Publ.,* **1**, No. 2, 1–8.

——, 1937: Theorie der aussertropischen Zyklonenbildung. *Meteor. Zeits.,* **54**, 186–190.

——, 1951: Extratropical cyclones. *Compendium of Meteorology,* T. F. Malone, Ed. American Meteorological Society, 577–598.

——, and H. Solberg, 1921: Meteorological conditions for the formulation of rain. *Geofys. Publ.,* **2**, No. 3, 1–60.

——, and H. Solberg, 1922: Life cycle of cyclones and the polar front theory of atmospheric circulation. *Geofys. Publ.,* **3**, No. 1, 1–18.

——, and C. Godske, 1936: On the theory of cyclone formation at extratropical fronts. *Astrophys. Norv.,* **1**, No. 6, 199–235.

——, and E. Palmén, 1937: Investigations of selected European cyclones by means of serial ascents. *Geofys. Publ.,* **12**, No. 2, 1–62.

——, and J. Holmboe, 1944: On the theory of cyclones. *J. Meteor.,* **1**, 1–22.

Bond, N. A., and R. G. Fleagle, 1985: Structure of a cold front over the ocean. *Quart. J. Roy. Meteor. Soc.,* **111**, 739–760.

Bosart, L. F., 1970: Mid-tropospheric frontogenesis. *Quart. J. Roy. Meteor. Soc.,* **96**, 442–471.

——, 1975: New England coastal frontogenesis. *Quart. J. Roy. Meteor. Soc.,* **97**, 457–482.

——, 1981: The Presidents' Day snowstorm of 18-19 February 1979: A subsynoptic scale event. *Mon. Wea. Rev.,* **109**, 1542–1566.

——, and S. C. Lin, 1984: A diagnostic analysis of the Presidents' Day storm of February 1979. *Mon. Wea. Rev.,* **112**, 2148–2177.

——, C. J. Vaudo and J. H. Helsdon, Jr., 1972: Coastal frontogenesis. *J. Appl. Meteor.,* **11**, 1236–1278.

Briggs, J., and W. T. Roach, 1963: Aircraft observations near jet streams. *Quart. J. Roy. Meteor. Soc.,* **89**, 225–247.

Brown, J. A., Jr., 1969a: A numerical investigation of hydrodynamic instability and energy conversions in the quasi-geostrophic atmosphere: Part I. *J. Atmos. Sci.,* **26**, 352–365.

——, 1969b: A numerical investigation of hydrodynamic instability and energy conversions in the quasi-geostrophic atmosphere: Part II. *J. Atmos. Sci.,* **26**, 366–375.

Browning, K. A., and T. W. Harrold, 1969: Air motion and precipitation growth in a wave depression. *Quart. J. Roy. Meteor. Soc.,* **95**, 288–309.

——, and ——, 1970: Air motion and precipitation at a cold front. *Quart. J. Roy. Meteor. Soc.,* **96**, 369–389.

——, and F. F. Hill, 1985: Mesoscale analysis of a polar trough interacting with a polar front. *Quart. J. Roy. Meteor. Soc.,* **111**, 445–462.

Businger, S., and R. J. Reed, 1989: Cyclogenesis in cold air masses. *Weather and Forecasting,* **4**, 133–156.

Buzzi, A., T. Nanni and M. Tagliazucca, 1977: Mid-tropospheric frontal zones: Numerical experiments with an isentropic coordinate primitive equation model. *Arch. Meteor. Geophys. Bioklim,* **A26**, 135–178.

——, A. Trevisan and G. Salustri, 1981: Internal frontogenesis: A two-dimensional model in isentropic semi-geostrophic coordinates. *Mon. Wea. Rev.,* **109**, 1053–1060.

Carbone, R. E., 1982: A severe winter squall line. Stormwide hydrodynamic structure. *J. Atmos. Sci.,* **39**, 258–279.

Carlson, T. N., 1980: Airflow through midlatitude cyclones and the comma cloud pattern. *Mon. Wea. Rev.,* **108**, 1498–1509.

Charney, J. G., 1947: The dynamics of long waves in a baroclinic westerly current. *J. Meteor.,* **4**, 135–162.

Chen, S.-J., and L. Dell'Osso, 1987: A numerical case study of East Asian coastal cyclogenesis. *Mon. Wea. Rev.,* **115**, 477–487.

Danard, M. B., and G. E. Ellenton, 1980: Physical influences on East Coast cyclogenesis. *Atmos.-Ocean,* **18**, 65–82.

Danielsen, E. F., 1964: Project Springfield Report. Defense Atomic Support Agency, Washington D.C. 20301, DASA 1517 (NTIS #AD-607980), 97 pp.

——, 1968: Stratospheric-tropospheric exchange based on radioactivity, ozone and potential vorticity. *J. Atmos. Sci.,* **25**, 502–518.

——, and V. A. Mohnen, 1977: Project dust storm report: ozone transport, in situ measurements and meteorological analyses of tropopause folding. *J. Geophys. Res.,* **82**, 5867–5878.

——, R. T. Hipskind, S. E. Gaines, G. W. Sachse, G. L. Gregory and G. F. Hill, 1987: Three-dimensional analysis of potential vorticity associated with tropopause folds and observed variations of ozone and carbon monoxide. *J. Geophys. Res.,* **92**, No. D2, 2103–2120.

Defant, A., 1912: Die Veränderungen in der allgemeinen Zirkulation der Atmosphäre in den gemässigten Breiten der Erde.

Wiener Akad. Wiss., Sitzungsberichte, **121**, 379-586. (Summary in *Meteor. Zeits.*, **30**, 126–138.)

Duncan, C. N., 1977: A numerical investigation of polar lows. *Quart. J. Roy. Meteor. Soc.*, **103**, 255–268.

Durran, D. R., and D. B. Weber, 1988: An investigation of the poleward edges of cirrus clouds associated with midlatitude jet streams. *Mon. Wea. Rev.*, **116**, 702–714.

Eady, E. T., 1949: Long waves and cyclone waves. *Tellus*, **1** (3), 33–52.

Edelmann, W., 1963: On the behavior of disturbances in a baroclinic channel. Sum. Rept. No. 2. Research in Objective Forecasting, Part F, Contract No. AF61 (052)–373, Deutscher Wetterdienst, Offenbach, 35 pp.

Eliassen, A., 1959: On the formation of fronts in the atmosphere. *The Atmosphere and the Sea in Motion*, B. Bolin, Ed. Rockefeller Institute Press, 277–287.

——, 1962: On the vertical circulation in frontal zones. *Geofys. Publ.*, **24**, 147–160.

——, 1966: Motions of intermediate scale: Fronts and cyclones. *Advances in Earth Science*, P. M. Hurley, Ed. MIT Press, 111–138.

——, and E. Kleinschmidt, 1957: Dynamic meteorology. *Handbuch der Physik*, **48**, S. Flügge, Ed. Springer-Verlag, 1–154.

Emanuel, K. A., 1983: On assessing local conditional symmetric instability from atmospheric soundings. *Mon. Wea. Rev.*, **111**, 2016–2033.

——, 1985: Frontal circulations in the presence of small moist symmetric instability. *J. Atmos. Sci.*, **42**, 1062–1071.

——, M. Fantini and A. J. Thorpe, 1987: Baroclinic instability in an environment of small stability to slantwise moist convection. Part I: Two-dimensional models. *J. Atmos. Sci.*, **44**, 1559–1573.

Farrell, B., 1982: The initial growth of disturbances in a baroclinic flow. *J. Atmos. Sci.*, **39**, 1663–1686.

——, 1984: Modal and non-modal baroclinic waves. *J. Atmos. Sci.*, **41**, 668–673.

——, 1985: Transient growth of damped baroclinic waves. *J. Atmos. Sci.*, **42**, 2718–2727.

——, 1989: Optimal excitation of baroclinic waves. *J. Atmos. Sci.*, **46**, 1193–1206.

Fjørtoft, R., 1955: On the use of space-smoothing in physical weather forecasting. *Tellus*, **7**, 462–480.

Gall, R., 1976a: A comparison of linear baroclinic instability theory with the eddy statistics of a general circulation model. *J. Atmos. Sci.*, **33**, 349–373.

——, 1976b: Structural changes of growing baroclinic waves. *J. Atmos. Sci.*, **33**, 374–390.

——, 1976c: The effects of released latent heat in growing baroclinic waves. *J. Atmos. Sci.*, **33**, 1686–1701.

——, R. T. Williams and T. L. Clark, 1987: On the minimum scale of surface fronts. *J. Atmos. Sci.*, **44**, 2562–2574.

Green, J. S. A., F. H. Ludlam and J. F. R. McIlveen, 1966: Isentropic relative-flow analysis and the parcel theory. *Quart. J. Roy. Meteor. Soc.*, **92**, 210–219.

Gyakum, J. R., 1983a: On the evolution of the QE II storm. I: Synoptic aspects. *Mon. Wea. Rev.*, **111**, 1137–1155.

——, 1983b: On the evolution of the QE II storm. II: Dynamic and thermodynamic structure. *Mon. Wea. Rev.*, **111**, 1156–1173.

Hayashi, Y., and D. E. Golder, 1981: The effects of condensational heating on midlatitude transient waves in their mature stage: Control experiments with a GFDL general circulation model. *J. Atmos. Sci.*, **38**, 2532–2539.

Heckley, W. A., and B. J. Hoskins, 1982: Baroclinic waves and frontogenesis in a non-uniform potential vorticity semi-geostrophic model. *J. Atmos. Sci.*, **39**, 1999–2016.

Hobbs, P. V., and P. O. G. Persson, 1982: The mesoscale and microscale structure and organization of clouds and precipitation in midlatitude cyclones. V: The substructure of narrow cold-frontal rainbands. *J. Atmos. Sci.*, **39**, 280–295.

Holton, J. R., 1979: *An Introduction to Dynamic Meteorology*, 2nd ed. Academic Press, 391 pp.

Hoskins, B. J., 1971: Atmospheric frontogenesis models: Some solutions. *Quart. J. Roy. Meteor. Soc.*, **97**, 139–153.

——, 1972: Non-Boussinesq effects and further development in a model of upper tropospheric frontogenesis. *Quart. J. Roy. Meteor. Soc.*, **98**, 532–547.

——, 1974: The role of potential vorticity in symmetric stability and instability. *Quart. J. Roy. Meteor. Soc.*, **100**, 480–482.

——, 1982: The mathematical theory of frontogenesis. *Annual Review of Fluid Mechanics*, **14**, Annual Reviews Inc., 131–151.

——, and F. P. Bretherton, 1972: Atmospheric frontogenesis models: Mathematical formulation and solution. *J. Atmos. Sci.*, **29**, 11–37.

——, I. Draghici and H. C. Davies, 1978: A new look at the ω-equation. *Quart. J. Roy. Meteor. Soc.*, **104**, 31–38.

——, and N. V. West, 1979: Baroclinic waves and frontogenesis. Part II: Uniform potential vorticity jet flows—cold and warm fronts. *J. Atmos. Sci.*, **36**, 1663–1680.

——, M. E. McIntyre and A. W. Robertson, 1985: On the use and significance of isentropic potential vorticity maps. *Quart. J. Roy. Meteor. Soc.*, **111**, 877–946.

Houze, R. A., Jr., and P. V. Hobbs, 1982: Organization and structure of precipitating cloud systems. *Advances in Geophysics*, **24**, 225–315.

Hsie, E.-Y., and R. A. Anthes, 1984: Simulations of frontogenesis in a moist atmosphere using alternative parameterizations of condensation and precipitation. *J. Atmos. Sci.*, **41**, 2701–2716.

——, ——, and D. Keyser, 1984: Numerical simulation of frontogenesis in a moist atmosphere. *J. Atmos. Sci.*, **41**, 2581–2594.

Keshishian, L. G., and L. F. Bosart, 1987: A case study of extended East Coast frontogenesis. *Mon. Wea. Rev.*, **115**, 100–117.

Keyser, D., 1986: Atmospheric fronts: an observational perspective. *Mesoscale Meteorology and Forecasting*, P. S. Ray, Ed. American Meteorological Society, Ch. 10.

——, and R. A. Anthes, 1982: The influence of planetary boundary layer physics in frontal structure in the Hoskins-Bretherton horizontal shear model. *J. Atmos. Sci.*, **39**, 1783–1802.

——, and M. J. Pecnick, 1985: A two-dimensional primitive equation model of frontogenesis forced by confluence and horizontal shear. *J. Atmos. Sci.*, **42**, 1259–1282.

——, and M. A. Shapiro, 1986: A review of the structure and dynamics of upper-level frontal zones. *Mon. Wea. Rev.*, **114**, 452–499.

Kleinschmidt, E., 1941: Zur theorie der labilen Anordnung. *Meteor. Zeits.*, **58**, 157–163.

——, 1957: Cyclones and anticyclones. Chap. IV in Eliassen and Kleinschmidt, 1957 (reference listed above).

Kotschin, N., 1932: Über die Stabilität von Margulesschen Diskontinuitäts-flächen. *Beitr. Phys. Atmos.*, **18**, 129–164.

Kuo, Y.-H., and R. J. Reed, 1988: Numerical simulation of an explosively deepening cyclone in the Eastern Pacific. *Mon. Wea. Rev.*, **116**, 2081-2105.

——, M. A. Shapiro and E. Donall, 1990: Interaction of baroclinic and diabatic processes in numerical simulations of a rapidly developing marine cyclone. *Mon. Wea. Rev.*, in press.

Kutzbach, G., 1979: *The Thermal Theory of Cyclones. A History of Meteorological Thought in the Nineteenth Century.* American Meteorological Society, 255 pp.

Mak, M., 1982: On moist quasi-geostrophic baroclinic instability. *J. Atmos. Sci.*, **39**, 2028-2037.

——, and P. R. Bannon, 1984: Frontogenesis in a moist semigeostrophic model. *J. Atmos. Sci.*, **41**, 3485-3500.

McGinnigle, J. B., M. V. Young and M. J. Baker, 1988: The development of instant occlusions in the North Atlantic. Met O 15 Internal Report No. 73, U.K. Meteorological Office, 25 pp.

McMurdie, L. A., and K. B. Katsaros, 1985: Atmospheric water distribution in a midlatitude cyclone observed by the Seasat Scanning Multichannel Microwave Radiometer. *Mon. Wea. Rev.*, **113**, 584-598.

Mechoso, C. R., and D. M. Sinton, 1983: On the energy analysis of the two-layer frontal model. *J. Atmos. Sci.*, **40**, 2069-2074.

Miller, J. E., 1948: On the concept of frontogenesis. *J. Meteor.*, **5**, 169-171.

Moore, G.W.K., 1987: Frontogenesis in a continuously varying potential vorticity fluid. *J. Atmos. Sci.*, **44**, 761-770.

——, and W. R. Peltier, 1987: Cyclogenesis in frontal zones. *J. Atmos. Sci.*, **44**, 384-409.

Mudrick, S. E., 1974: A numerical study of frontogenesis. *J. Atmos. Sci.*, **31**, 869-892.

Mullen, S. L., 1979: An investigation of small synoptic scale cyclones in polar air streams. *Mon. Wea. Rev.*, **107**, 1636-1647.

——, 1983: Explosive cyclogenesis associated with cyclones in polar air streams. *Mon. Wea. Rev.*, **111**, 1537-1553.

Namias, J., and P. F. Clapp, 1949: Confluence theory of the high tropospheric jet stream. *J. Meteor.*, **6**, 330-336.

Newton, C. W., 1954: Frontogenesis and frontolysis as a three-dimensional process. *J. Meteor.*, **11**, 449-461.

——, and A. Trevisan, 1984a: Clinogenesis and frontogenesis in jet-stream waves. Part I: Analytical relations to wave structure. *J. Atmos. Sci.*, **41**, 2717-2734.

——, and ——, 1984b: Clinogenesis and frontogenesis in jet-stream waves. Part II: Channel model numerical experiments. *J. Atmos. Sci.*, **41**, 2735-2755.

Orlanski, I., 1968: Instability of frontal waves. *J. Atmos. Sci.*, **25**, 178-200.

——, and B. B. Ross, 1984: The evolution of an observed cold front. Part II: Mesoscale dynamics. *J. Atmos. Sci.*, **41**, 1669-1703.

——, ——, L. Palinsky and R. Shaginaw, 1985: Advances in the theory of atmospheric fronts. *Advances in Geophysics*, **27**, 223-252.

Palmén, E., 1951: The role of atmospheric disturbances in the general circulation (Symons Memorial Lecture). *Quart. J. Roy. Meteor. Soc.*, **77**, 337-354.

——, and K. M. Nagler, 1948: An analysis of the wind and temperature distribution in the free atmosphere over North America in a case of approximately westerly flow. *J. Meteor.*, **5**, 58-64.

——, and ——, 1949: The formation and structure of a large-scale disturbance in the westerlies. *J. Meteor.*, **6**, 227-242.

Pedlosky, J., 1964a: The stability of currents in the atmosphere and ocean: Part I. *J. Atmos. Sci.*, **21**, 201-219.

——, 1964b: The stability of currents in the atmosphere and ocean: Part II. *J. Atmos. Sci.*, **21**, 342-353.

Petersen, R. A., L. W. Uccellini, A. Mostek and D. A. Keyser, 1984: Delineating mid- and low-level water vapor patterns in pre-convective environments using VAS moisture channels. *Mon. Wea. Rev.*, **112**, 2178-2198.

Petterssen, S., 1936: Contribution to the theory of frontogenesis. *Geofys. Publ.*, **11**, No. 6, 27 pp.

——, 1956: *Weather Analysis and Forecasting*, 2nd ed., Vol. 1. McGraw-Hill, 428 pp.

——, and J. M. Austin, 1942: Fronts and frontogenesis in relation to vorticity. *Papers in Physical Oceanography and Meteorology*, VII, No. 2. MIT and Woods Hole Oceanographic Institution, 37 pp.

——, G. E. Dunn and L. L. Means, 1955: Report of an experiment in forecasting of cyclone development. *J. Meteor.*, **12**, 58-67.

——, D. L. Bradbury and K. Pedersen, 1962: The Norwegian cyclone models in relation to heat and cold sources. *Geofys. Publ.*, **24**, 243-280.

Phillips, N. A., 1956: The general circulation of the atmosphere: A numerical experiment. *Quart. J. Roy. Meteor. Soc.*, **82**, 124-164.

Reed, R. J., 1955: A study of a characteristic type of upper-level frontogenesis. *J. Meteor.*, **12**, 542-552.

——, 1979: Cyclogenesis in polar air streams. *Mon. Wea. Rev.*, **107**, 38-52.

——, and F. Sanders, 1953: An investigation of the development of a mid-tropospheric frontal zone and its associated vorticity field. *J. Meteor.*, **10**, 338-349.

——, and M. D. Albright, 1986: A case study of explosive cyclogenesis in the Eastern Pacific. *Mon. Wea. Rev.*, **112**, 2297-2319.

——, and W. Blier, 1986a: A case study of comma cloud development in the Eastern Pacific. *Mon. Wea. Rev.*, **114**, 1681-1695.

——, and ——, 1986b: A further case study of comma cloud development in the Eastern Pacific. *Mon. Wea. Rev.*, **114**, 1696-1708.

——, A. J. Simmons, M. D. Albright and P. Undén, 1988: The role of latent heat release in explosive cyclogenesis: Three examples based on ECMWF Operational Forecasts. *Weather and Forecasting*, **3**, 217-229.

Rogers, E., and L. F. Bosart, 1986: An investigation of explosively deepening oceanic cyclones. *Mon. Wea. Rev.*, **114**, 702-718.

Ross, B. B., and I. Orlanski, 1978: The circulation associated with a cold front. Part II: Moist case. *J. Atmos. Sci.*, **35**, 445-465.

——, and ——, 1982: The evolution of an observed front. Part I: Numerical simulation. *J. Atmos. Sci.*, **39**, 296-326.

Sanders, F., 1955: An investigation of the structure and dynamics of an intense surface frontal zone. *J. Meteor.*, **12**, 542-552.

——, 1987: Skill of NMC operational dynamical models in prediction of explosive cyclogenesis. *Weather and Forecasting*, **2**, 322-336.

——, and J. R. Gyakum, 1980: Synoptic-dynamic climatology of the "bomb." *Mon. Wea. Rev.*, **108**, 1589-1606.

——, and L. F. Bosart, 1985: Mesoscale structure in the megalopolitan snowstorm of 11-12 February 1983. Part I: Frontogenetical forcing and symmetric instability. *J. Atmos. Sci.*, **42**, 1050–1061.

Sawyer, J. S., 1949: The significance of dynamic instability in atmospheric motions. *Quart. J. Roy. Meteor. Soc.*, **75**, 364–374.

——, 1956: The vertical circulation at meteorological fronts and its relation to frontogenesis. *Proc. Roy. Soc. London*, **A234**, 346–362.

Shapiro, M. A., 1975: Simulation of upper-level frontogenesis with a 20-level isentropic coordinate primitive equation model. *Mon. Wea. Rev.*, **103**, 591–604.

——, 1976: The role of turbulent heat flux in the generation of potential vorticity in the vicinity of upper-level jet stream systems. *Mon. Wea. Rev.*, **104**, 892–900.

——, 1978: Further evidence of the mesoscale and turbulent structure of upper level jet stream-frontal zone systems. *Mon. Wea. Rev.*, **106**, 1100–1111.

——, 1980: Turbulent mixing within tropopause folds as a mechanism for the exchange of chemical constituents between the stratosphere and the troposphere. *J. Atmos. Sci.*, **37**, 994–1004.

——, 1981: Frontogenesis and geostrophically forced secondary circulations in the vicinity of jet stream-frontal zone systems. *J. Atmos. Sci.*, **38**, 954–973.

——, 1983: Mesoscale weather systems of the central United States. The National STORM program: Scientific and Technological Bases and Major Objectives, R. A. Anthes, Ed. University Corporation for Atmospheric Research, Boulder, CO 80307-3000, 3.1–3.77.

——, 1984: Meteorological tower measurements of a surface cold front. *Mon. Wea. Rev.*, **112**, 1634–1639.

——, T. Hampel, D. Rotzoll and F. Mosher, 1985: The frontal hydraulic head: A micro-α scale ($\sim$1 km) triggering mechanism of mesoconvective weather systems. *Mon. Wea. Rev.*, **113**, 1166–1183.

Simmons, A. J., and B. J. Hoskins, 1976: Baroclinic instability on the sphere: Normal modes of the primitive and quasi-geostrophic equations. *J. Atmos. Sci.*, **33**, 1454–1477.

——, and ——, 1978: The life cycles of some nonlinear baroclinic waves. *J. Atmos. Sci.*, **35**, 414–432.

——, and ——, 1979: The downstream and upstream development of unstable baroclinic waves. *J. Atmos. Sci.*, **35**, 1239–1254.

Simons, T. J., 1972: The nonlinear dynamics of cyclone waves. *J. Atmos. Sci.*, **29**, 38–52.

Sinton, D. M., and C. R. Mechoso, 1984: Nonlinear evolution of frontal waves. *J. Atmos. Sci.*, **41**, 3501–3517.

Solberg, H., 1928: Integrationen der atmosphärischen Störungsgleichungen. *Geofys. Publ.*, **5**, No. 9, 1–120.

Song, R. T., 1971: A numerical study of the three-dimensional structure and energetics of unstable disturbances in zonal currents: Part I. *J. Atmos. Sci.*, **28**, 549–564.

Staff Members, University of Chicago, 1947: On the general circulation of the atmosphere in middle latitudes. *Bull. Amer. Meteor. Soc.*, **28**, 255–280.

Staley, D. O., 1960: Evaluation of potential vorticity changes near the tropopause and the related vertical motions, vertical advection of vorticity and transfer of radioactive debris from stratosphere to troposphere. *J. Meteor.*, **17**, 591–620.

——, and R. L. Gall, 1977: On the wavelength of maximum baroclinic instability. *J. Atmos. Sci.*, **34**, 1669–1688.

Stauffer, D. R., and T. T. Warner, 1987: A numerical study of cold-air damming and coastal frontogenesis. *Mon. Wea. Rev.*, **115**, 799–821.

Stone, P. H., 1966: Frontogenesis by horizontal wind deformation fields. *J. Atmos. Sci.*, **23**, 455–465.

Sutcliffe, R. C., 1947: A contribution to the problem of development. *Quart. J. Roy. Meteor. Soc.*, **73**, 370–383.

Takayabu, I., 1986: Roles of the horizontal advection on the formation of surface fronts and on the occlusion of a cyclone developing in the baroclinic easterly jet. *J. Meteor. Soc. Japan*, **64**, 329–345.

Thorpe, A. J., and K. A. Emanuel, 1985: Frontogenesis in the presence of small stability to slantwise convection. *J. Atmos. Sci.*, **42**, 1809–1824.

Todsen, M., 1964: A study of the vertical circulations in a cold front. Part IV of Final Report, Air Force Cambridge Laboratories, OAR Contr. No. AF61 (052)–525.

Trenberth, K. E., 1978: On the interpretation of the diagnostic quasi-geostrophic omega equation. *Mon. Wea. Rev.*, **106**, 131–137.

Uccellini, L. W., 1986: The possible influence of upstream upper-level baroclinic processes on the development of the QE II storm. *Mon. Wea. Rev.*, **114**, 1019–1027.

——, P. J. Kocin, R. A. Petersen, C. H. Wash and K. F. Brill, 1984: The Presidents' Day cyclone of 18-19 February 1979: Synoptic overview and analysis of the subtropical jet streak influencing the precyclogenetic period. *Mon. Wea. Rev.*, **113**, 31–55.

——, D. Keyser, K. F. Brill and C. H. Wash, 1985: Presidents' Day cyclone of 18-19 February 1979: Influence of upstream trough amplification and associated tropopause folding on rapid cyclogenesis. *Mon. Wea. Rev.*, **112**, 962–988.

——, R. A. Petersen, K. F. Brill, P. J. Kocin and J. J. Toccillo, 1987: Synergistic interactions between an upper level jet streak and diabatic processes that influence the development of a low level jet and a secondary coastal cyclone. *Mon. Wea. Rev.*, **115**, 2227–2261.

Wallace, J. M., and P. V. Hobbs, 1977: *Atmospheric Science: An Introductory Survey.* Academic Press, 417 pp.

Welander, P., 1963: Steady plane fronts in a rotating fluid. *Tellus*, **15**, 33–43.

Williams, R. T., 1967: Atmospheric frontogenesis: A numerical experiment. *J. Atmos. Sci.*, **24**, 627–641.

——, 1972: Quasi-geostrophic versus non-geostrophic frontogenesis. *J. Atmos. Sci.*, **29**, 3–10.

——, 1974: Numerical simulation of steady-state fronts. *J. Atmos. Sci.*, **31**, 1286–1296.

——, L. C. Chou and C. J. Cornelius, 1981: Effects of condensation and surface motion on the structure of steady-state fronts. *J. Atmos. Sci.*, **38**, 2365–2376.

WMO, 1973: *The Use of Satellite Pictures in Weather Analysis and Forecasting.* Tech. Note No. 124, WMO No. 333. World Meteorological Organization, Geneva, 275 pp.