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ISBN: 978-0-12-088542-8
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Academic Press
The Mesoscale Structure of Extratropical Cyclones and Middle and High Clouds

10.1. INTRODUCTION

Extratropical cyclones have been found to account for about half of the warm season precipitation over the United States (Heideman and Fritsch, 1988), and the warm conveyor belts associated with these cyclones produce approximately half of the wintertime precipitation in middle and high latitudes (Eckhardt et al., 2004). Thus these storms, also referred to as mid-latitude cyclones, play an important role in the hydrological cycle of the mid-latitudes. In Chapter 9 we considered how the dynamical behavior of an atmospheric circulation system can be characterized by the scaling parameter \( \lambda R \), the Rossby radius of deformation [Eq. (9.3)]. A substitution of typical values of the parameters in Eq. (9.3) for tropical and extratropical cyclones (averaged over the entire cyclone) shows that for the largest scale of these cyclonic disturbances, the atmosphere behaves as a two-dimensional, quasi-balanced fluid (Table 10.1). However an enormous spatial variation of \( \lambda R \) exists within extratropical cyclones and thus a variety of clouds and mesoscale precipitating phenomena are embedded within these systems. As a result, extratropical cyclones also participate in the global energy balance. Finally, these storms may also produce a wide range of severe weather. In an analysis of what has been referred to as The Storm of the Century, Kocin et al. (1995) described the widespread heavy snowfall, coastal flooding, squall lines, thunderstorms and tornadoes associated with an extratropical cyclone that developed and progressed along the southern and eastern coasts of the United States between the 12 and 14 March, 1993. Debate continues as to whether the mesoscale phenomena associated with mid-latitude cyclones are simply a consequence of the large-scale motions of the cyclone, or whether they may modulate the cyclone through upscale forcing. Zhang and Harvey (1995), for example, document a case in which a squall line enhanced the large-scale baroclinic environment, making it more favorable for subsequent surface cyclogenesis. The focus in this chapter is on the processes and properties of various mesoscale features generated by extratropical cyclones.
Synoptic-scale forcing and processes are only discussed when relevant to the mesoscale features of interest.

10.2. LARGE-SCALE PROCESSES THAT DETERMINE MESOSCALE FEATURES

Mesoscale cloud and precipitation features are initiated by two mechanisms—forcing on the mesoscale by inhomogeneities in the surface (such as terrain features) and instabilities in the larger scale environment. Terrain-forced features are discussed in Chapter 11. Here we discuss the large-scale processes that produce an environment that is stable or unstable with respect to mesoscale cloud and precipitation systems. We have seen that the type of cloud and precipitation of a system is determined by six factors: (1) water vapor content of the air (both relative and absolute humidity), (2) temperature, (3) aerosol types and amounts, (4) static stability, (5) vertical motion, and (6) vertical shear of the horizontal wind. As these atmospheric properties vary greatly throughout extratropical cyclones, these storms contain a rich variety of clouds and mesoscale precipitating systems. After a brief description of the physical processes determining these parameters on large scales of motion (100 km and greater), we discuss their variation in extratropical cyclones, which in turn affects the variation of clouds and precipitation in these systems.

10.2.1. Water Vapor Content

Chapters 2 and 4 describe in detail the equations that determine the temporal variation of water vapor as well as liquid water and ice. For convenience, we repeat the continuity equation for the water vapor mixing ratio appropriate for large-scale models,

$$\frac{\partial r_v}{\partial t} = -V \cdot \nabla r_v - w \frac{\partial r_v}{\partial z} + E - C + F_{rv},$$  \hspace{1cm} (10.1)
where $E$ represents evaporation (from the surface or from precipitation), $C$ represents condensation (including sublimation), $F_{rv}$ represents unresolvable (subgrid-scale) transports, $V$ is the horizontal velocity vector, and $\nabla$ is the horizontal gradient operator. Over short periods of time and for the large scales of motion considered here, horizontal and vertical transports are the major processes contributing to local changes in water vapor. For longer time periods (greater than 12 h), evaporation from the surface is important over water and over land surfaces that are moister and warmer than the air immediately above the surface. Condensation is a key removal mechanism in precipitating systems, whereas evaporation of cloud water and precipitation can be locally significant over short time periods. Subgrid-scale (turbulent) transports are greatest in the unstable planetary boundary layer where water vapor evaporated from the surface is transported upward.

10.2.2. Temperature

Detailed forms of the thermodynamic equation are presented in Chapter 2. A simplified form appropriate for interpreting the temporal variation of the large-scale temperature is

$$\frac{\partial \theta}{\partial t} = -V \cdot \nabla \theta - w \frac{\partial \theta}{\partial z} + \frac{\theta}{c_p T} Q + F_\theta,$$

(10.2)

where $\theta$ is potential temperature and $Q$ is the net grid scale averaged diabatic heating. All of the other variables have been defined previously. The first term of the right-hand side represents horizontal advection of potential temperature, while the second term is the vertical advection of potential temperature, which in a statically stable atmosphere produces cooling with upward motion and warming with subsiding motion. The diabatic heating term $Q$ includes latent heating and cooling effects associated with condensation or evaporation, as well as radiation. In precipitation systems, there is a close balance between adiabatic cooling associated with vertical advection and diabatic heating. The last term $F_\theta$ is the subgrid-scale transport of heat and includes sensible heating from the surface and the upward transport in the PBL. On the large scale, above the PBL and in the absence of precipitation, horizontal and vertical advection are the largest terms. During the daytime in the heated PBL, the last term, dominates.

10.2.3. Aerosol Types and Amounts

Condensation does not usually occur at relative humidities of exactly 100% (Chapter 4). The presence and type of cloud depend on the amount and distribution of aerosols in the atmosphere, in particular, cloud condensation nuclei and ice nuclei. The effects of these aerosols are discussed in Chapter 4 and are mentioned here for completeness. They are not presently explicitly represented in most large-scale models, however some cloud resolving models
now include explicit aerosol schemes (e.g. Saleeby and Cotton, 2004; Ekman et al., 2004).

10.2.4. Static Stability

The evolution of cloud and precipitation systems depends on the mean vertical motion and the atmospheric static stability. Static stability is described in detail in Chapter 2. As described in this chapter, conditional instability refers to the state of a parcel of air as it is lifted through its environment. Under such conditions a parcel of air at the environmental temperature is stable to all vertical displacements if it is unsaturated, but is unstable to upward vertical displacements if it is saturated, and is unstable to downward vertical displacements if it is saturated and contains cloud water. Convective instability, in contrast, refers to the state of instability when lifting an entire layer of air. If such a layer is lifted until it becomes completely saturated, then the layer will become unstable. The condition for convective instability, which is sometimes referred to as potential instability, is that equivalent potential temperature decreases throughout the layer,

$$\frac{\partial \theta_e}{\partial z} < 0.$$  (10.3)

It is possible for a layer of air to be convectively unstable but conditionally stable. Convective instability is often associated with the large-scale environment prior to the development of severe convective storms and tornadoes, when warm, dry air overlies warm, moist air, with an associated rapid decrease of $\theta_e$ with height. If the layer is lifted, it rapidly becomes unstable and favorable for severe thunderstorm development. When the mean vertical motion is near zero and the atmosphere is conditionally stable, fogs or layered clouds occur (Chapter 6). When the mean vertical velocity is upward (typically a few centimeters per second) and the atmosphere is conditionally stable, deep layers of nonconvective (stratiform) clouds are produced. Under conditionally unstable conditions, convective clouds and precipitation can occur, even with near-zero mean (large-scale) vertical velocities or weak subsidence.

If we define a static stability parameter $\gamma_0 \equiv \frac{\partial \theta}{\partial z}$, a simple equation describing the temporal variation of the large-scale static stability can be derived from Eq. (10.2):

$$\frac{\partial \gamma_0}{\partial t} = -\nabla \cdot \nabla \theta - \frac{\partial w \gamma_0}{\partial z} + \frac{\partial}{\partial z} \left( \frac{\theta}{c_p T Q} \right) + \frac{\partial F_\theta}{\partial z}.$$  (10.4)

The first term on the right-hand side of Eq. (10.4) represents the variation of the horizontal advection of potential temperature with height; for example, cold advection overlying warm advection contributes to destabilization. The second term represents the effect of vertical stretching of a column, i.e. if $\gamma_0$ is constant in the vertical, an increase of upward motion with height (stretching) represents
destabilization. This process is effective in producing or destroying temperature inversions. The third term represents the effect of differential heating in the vertical. For example, a decrease in diabatic heating with height, as occurs above the region of maximum latent heating associated with cumulus convection, destabilizes the environment. Another example of this process is radiative cooling near the tops of layered clouds which destabilizes this region. The final term represents the vertical variation of turbulent heat fluxes and is largest in the heated PBL.

Conditional and convective instabilities do not consider the effects of rotation, but rotation affects the stability of fluid motions. Bennetts and Hoskins (1979) and Emanuel (1979, 1982, 1983a,b) discuss the combined influence of rotation and static stability in a theory of conditional symmetric instability (CSI) in which the atmosphere is convectively stable and inertially stable, and yet is unstable to slantwise ascent. We start by examining dry symmetric instability. As just stated, the condition for dry absolute instability is that the potential temperature decreases with height \( \partial \theta / \partial z < 0 \). The condition for inviscid, inertial instability in the Northern Hemisphere is \( \partial M_g / \partial x < 0 \), where \( M_g = v_g + fx \) is the geostrophic absolute momentum of a geostrophically balanced mean state, \( v_g \) is the geostrophic wind in the direction perpendicular to the temperature gradient (and is in thermal wind balance), \( x \) is the distance in the direction of the temperature gradient (perpendicular to the thermal wind) with \( x \) increasing toward the warmer air, and \( f \) is the Coriolis parameter. In the absence of friction, \( M_g \) is approximately conserved. A parcel may be absolutely stable to vertical displacement \( \partial \theta / \partial z > 0 \), and inertially stable to horizontal displacements \( \partial M_g / \partial x > 0 \), but unstable with respect to slantwise displacements by dry symmetric instability. The condition for dry symmetric instability is that the \( M_g \) surfaces slope less steeply than the isentropic surfaces. Dry symmetric instability can therefore be viewed either as dry absolute instability on an \( M_g \) surface, or inertial instability on an isentropic surface. Any slantwise displacement that occurs at an angle between the slopes of the \( M_g \) and isentropic surfaces will therefore release the symmetric instability.

For moist slantwise convection, CSI occurs at each height where the environmental lapse rate along an \( M_g \) surface lies between the dry- and saturated-adiabatic lapse rates, and hence is conditionally unstable along an \( M_g \) surface. The CSI condition is then one in which the saturation equivalent potential temperature decreases with height along an \( M_g \) surface. Thus if a parcel moves upward along a surface of constant \( M_g \) and becomes warmer than its environment due to the release of latent heat, the atmosphere is in a state of conditional symmetric instability. A conditionally stable atmosphere may possess conditional symmetric instability. Convection arising from the release of CSI is called moist slantwise convection. The release of CSI is thought to be an important process in producing rainbands in extratropical cyclones. In environments near the frontal zones of extratropical cyclones a variety of instability types may be found, ranging from convectively unstable air in the
warm sector, to CSI north of the frontal boundary, and weak symmetric stability to the north of that.

Finally, as described by Schultz and Schumacher (1999) in a review article on the use and misuse of CSI as a diagnostic tool, equivalent potential temperature instead of saturation equivalent potential temperature is often, although incorrectly, used to assess CSI. When using equivalent potential temperature, it is convective or potential symmetric instability (PSI), rather than conditional symmetric instability, that is being assessed. As there can be significant differences between equivalent potential temperature and saturated equivalent potential temperature, there could be a significant difference between conditional and convective symmetric instabilities. Thus the term CSI should only be employed when using saturated equivalent potential temperature, while PSI should be utilized when using equivalent potential temperature. However, Moore et al. (2005) make the point that the instability is not realized unless the atmosphere becomes saturated in the presence of large-scale lifting, at which time the equivalent potential temperature and the saturated equivalent potential temperature are equal. It has been found to be more convenient operationally to use the equivalent potential temperature, and require that the relative humidity be at least 80%, in order to diagnose regions that are susceptible to CSI (Market and Cissell, 2002).

10.2.5. Vertical Motion

Along with moisture content and static stability, vertical motion is one of the most important properties of the large-scale environment that determines the presence and type of clouds and precipitation systems. Large-scale upward motion favors cloud and precipitation development because it both cools and destabilizes the air. In addition, the low-level convergence associated with rising motion in the middle troposphere is associated with moisture convergence in all but the driest air masses. By contrast, sinking air becomes more stable, relative humidity decreases, and low-level moisture divergence usually results.

A useful diagnostic equation for isolating the large-scale physical processes associated with vertical motion is the quasi-geostrophic omega equation, derived from the vorticity equation and the first law of thermodynamics using the assumption that the vorticity and the large-scale wind are in quasi-geostrophic balance (see Holton (1979) for a derivation). A convenient form of the omega equation with pressure as the vertical coordinate can be written as

\[
\nabla^2 \omega + \frac{f_0^2}{\sigma} \frac{\partial^2 \omega}{\partial p^2} = \frac{f}{\sigma} \frac{\partial}{\partial p} [V_g \cdot \nabla (\zeta + f)] + \frac{R}{P \sigma} \nabla^2 V_g \cdot \nabla T - \frac{R}{c_p \sigma} \nabla^2 Q - \frac{f_0}{\sigma} \frac{\partial F_\zeta}{\partial p},
\]

(10.5)
where the static stability parameter $\sigma$ is given by

$$\sigma = -(1/\rho \theta)(\partial \theta / \partial p),$$

and $F_\zeta$ represents the contribution of subgrid-scale effects (friction) to the temporal rate of change of the vertical component of the relative vorticity ($\zeta$).

To interpret Eq. (10.5), wave forms may be assumed for the horizontal and vertical variation of $\omega$ (Holton, 1979),

$$\omega \approx \sin[\pi (p/p_0)] \sin(kx) \sin(ly),$$

where $k$ and $l$ are horizontal wave numbers and $p_0$ is a reference pressure. Using the assumptions made in Eq. (10.6), the left-hand side of Eq. (10.5) can be written as

$$\left(\nabla^2 + f_0^2 \frac{\sigma}{\partial p^2}\right) \omega \approx -\left[(k^2 + l^2) + \frac{1}{\sigma} \left(\frac{f_0}{p_0}\right)^2\right] \omega. \tag{10.7}$$

Equation (10.7) demonstrates that the left-hand side of Eq. (10.5) is proportional to $-\omega$, and hence that the left-hand side of Eq. (10.5) is proportional to $w$. Thus, the Omega equation, Eq. (10.5), can be interpreted as follows. The first term on the right-hand hand side represents the vertical derivative of the horizontal vorticity advection. In most situations in the atmosphere, the vorticity advection is much smaller in the lower troposphere than in the middle to upper troposphere, and the sign of this term is thus generally determined by the horizontal vorticity advection. Therefore, rising (sinking) vertical motion is generally proportional to positive (negative) vorticity advection in the Northern Hemisphere. The second term on the right-hand side represents the Laplacian of the horizontal temperature advection, and thus $-\omega \propto V_g \cdot \nabla T$. In regions of warm air (cold air) advection, this term is positive (negative) and contributes to rising (sinking) motion. The third term on the right-hand side is the Laplacian of diabatic heating and thus $-\omega \propto Q$ (as the sign on this term is negative). Regions of maximum diabatic heating are therefore associated with upward motion.

The last term in Eq. (10.5) represents the effect of subgrid-scale motions (turbulence). In the PBL, its effect may be estimated by assuming a quadratic stress law for the frictional terms in the equations of motion,

$$\frac{\partial u}{\partial t} = \cdots - C_D |V| u/h,$$

$$\frac{\partial v}{\partial t} = \cdots - C_D |V| v/h, \tag{10.8}$$

where $C_D$ is the drag coefficient, $u$ and $v$ are the mean horizontal wind components in the PBL, and $h$ is the depth of the PBL. With the frictional terms
represented by Eq. (10.11), the linearized form of $F_\zeta$ is:

$$F_\zeta(p_s) \approx -K \zeta(p_s), \quad (10.9)$$

where $K$ is a mean value of $C_D|\mathbf{V}|/h$. Thus, using Eqs (10.5), (10.7) and (10.9), the vertical velocity near the top of the PBL is approximately

$$\omega \propto \frac{\partial F_\zeta}{\partial p} \propto F_\zeta(p_S), \quad (10.10)$$

where we have used the fact that $F_\zeta$ vanishes near the top of the PBL. From (10.9) and (10.10) we see that surface friction induces upward motion in cyclonic systems and downward motion in anticyclonic systems. This effect is sometimes called *Ekman pumping* or *Ekman suction* depending on its sign.

10.2.6. Vertical Shear of the Horizontal Wind

A fifth property of the environment important in the development of some types of clouds and convective systems is the vertical shear of the horizontal wind. In the PBL, the wind shear organizes fair-weather cumulus clouds into bands, rolls, rings, and streets (Chapter 7). In addition, strong vertical wind shear is a major factor in determining the organization and structure of cumulonimbus clouds. Wind shear affects the entrainment rate, the strength, movement, precipitation efficiency, and lifetime of convective clouds and storms (Chapter 8). Wind shear is also a factor in the splitting of severe thunderstorms and the development of rotating storms and hail and tornado-producing thunderstorms.

The development of wind shear in the large-scale environment is closely tied to the development of baroclinicity, since the thermal wind balance is approximately satisfied for these scales of motion. Figure 10.1 shows a horizontal cross section of a strong baroclinic zone and the associated wind shear (Shapiro et al., 1984). Baroclinicity on these scales is produced primarily by two mechanisms - frontogenetic processes and differential heating associated with latent heat release. Frontogenetic processes are important in extratropical cyclone systems. Confluence and deformation in the large-scale environment and an ageostrophic response of the atmosphere to thermal-wind imbalances produced by the changing baroclinicity are key elements of the frontogenesis process. A local increase in the horizontal temperature gradient through deformation of the wind, disrupts thermal wind balance as the horizontal temperature gradient becomes too large for the associated vertical wind shear. The atmosphere produces a thermally-direct ageostrophic circulation transverse to the baroclinic zone, thereby re-establishing thermal wind balance (Koch, 1984; Sanders and Bosart, 1985; Keyser and Shapiro, 1986). Frontogenesis therefore enhances vertical motion through this thermally-direct ageostrophic circulation. Figure 10.2 shows the transverse ageostrophic circulation associated with frontogenesis in a numerical model. This vertical circulation, in addition
to playing an essential role in the frontogenesis process and the development of wind shear, destabilizes the environment on the warm, moist side of fronts and triggers clouds and precipitation systems. Baroclinicity can also be produced by differential heating associated with the release of latent heat.

Another mechanism for producing low-level wind shear is surface friction. The general decrease of frictional effects with height in the lower troposphere results in vertical wind shear. For example, the diurnal variation in the depth of the PBL and the intensity of turbulent mixing of momentum can lead to the development of low-level jets (Blackadar, 1957; Bonner, 1966, 1968). The horizontal convergence and wind shear associated with these jets can significantly affect the development of convective storms.
FIGURE 10.2  Cross section of transverse ageostrophic circulation \((v_{ag}, w)\) and potential temperature (dashed lines, contour interval 5 K) after a 24-h integration of a two-dimensional primitive-equation model of frontogenesis (Keyser and Pecnick, 1985a,b) due to confluence in the presence of advection. Location of upper-level jet in along-front-velocity component is indicated by J; magnitudes of components of transverse ageostrophic circulation are represented by vector scales on lower right margins of figure. (From Keyser and Shapiro (1986))

10.2.7. Ocean versus Land Extratropical Cyclones

The conceptual model of extratropical cyclone development discussed below is modified somewhat for cyclones over the ocean. Two major differences over the ocean are the greater availability of moisture and the smoother, more homogeneous surface. The latter allows more rapid development and ultimately more intense cyclones. Rapid development can be pronounced when the sea surface is relatively warm, thus destabilizing lower levels and leading to explosively deepening cyclones, often referred to as “bombs” (Sanders and Gyakum, 1980). In addition, the more homogeneous surface favors the development of organized mesoscale convective clouds in the PBL (Chapter 7).

10.3. MESOSCALE STRUCTURE OF EXTRATROPICAL CYCLONES

10.3.1. Introduction

Driven inexorably by differential radiative heating between high and low latitudes, the middle-latitude atmosphere is characterized by large-scale horizontal temperature gradients and, through the thermal wind relationship,
westerly winds that normally increase with height throughout the troposphere. The poleward decrease of temperature is rarely uniform, and, instead, is usually concentrated in relatively narrow baroclinic zones or fronts. These baroclinic zones become unstable with respect to wavelike perturbations (Charney, 1947; Eady, 1949), and the result is the development of cyclones in the baroclinic zone. The wavelength of maximum instability depends on the static stability and horizontal temperature gradient (Staley and Gall, 1977). On the average it is around 3000 km.

As cyclones develop, cold air is carried southward to the rear of the cyclone while warm air is carried northward ahead of the cyclone. For convenience, we refer to the Northern Hemisphere in these discussions. Confluence and deformation associated with the developing circulation produce increasing horizontal temperature gradients in narrow bands, i.e. warm and cold fronts. Temperature changes associated with horizontal advection and vertical motion destroy thermal wind balance, and the resulting ageostrophic motions produce organized regions of divergence, convergence, and associated vertical motion. Extratropical cyclones dominate the large-scale variability of the weather of middle latitudes. Figure 10.3 shows visible satellite imagery of a mature cyclone over North America. The circulation of this cyclone covers a significant portion of North America and adjacent waters east of the Rocky Mountains.

Many aspects of the mesoscale structure of extratropical cyclones, discussed in detail in this text, include fogs and stratocumulus clouds (Chapter 6), cumulus clouds (Chapter 7), severe thunderstorms and tornadoes (Chapter 8), squall lines and other mesoscale convective systems (Chapter 9), and middle- and high-level
10.3.2. Conceptual Models of Extratropical Cyclones

While the focus of this chapter is on the mesoscale aspects of extratropical cyclones, it is useful to briefly describe the synoptic-scale organization of these systems, as the synoptic-scale features not only influence the mesoscale aspects, but are also influenced by the mesoscale features of these storms. Since its development during the early part of the 20th century, the Norwegian Cyclone model (Bjerknes, 1919; Bjerknes and Solberg, 1921, 1922) has formed the basis of our understanding of the life cycle of extratropical cyclones, an amazing achievement given that this conceptual model was developed during a time period of highly limited upper air data. The Norwegian Cyclone model is shown in Fig. 10.4. The development of the frontal wave from its incipient phase, through the mature cyclogenesis phase, and finally terminating in a frontal occlusion phase are all demonstrated. At the northern tip of the occluded front, a seclusion of warm air from the warm sector occurs as a result of being trapped during the occlusion process. The Norwegian Cyclone model also includes a description of the vertical structure of the warm, cold and occluded fronts, and the precipitation and cloud development associated with these fronts (bottom portion of Fig. 10.4), and Fig. 10.5 illustrates the typical cloud types in various
regions of the cyclone. This model, sometimes referred to as the ideal or classical cyclone model, was developed over 90 years ago and remains a useful, although somewhat simplified conceptual model.

Research conducted after World War II began to demonstrate that there may be more to the dynamic and microphysical aspects of extratropical cyclones than was suggested by the Norwegian Cyclone model. Observational studies indicated that the cyclone and frontal structures did not always extend continuously from the surface to the upper atmosphere, and that different dynamics were responsible for upper and lower-level fronts; early baroclinic wave simulations (e.g. Hoskins, 1976) produced frontal characteristics that had not as yet been observed. Based on previous modeling and observational studies, as well as observations from three field experiments investigating marine cyclone development, Shapiro and Keyser (1990) proposed several significant modifications to the Norwegian Cyclone model.

In the Shapiro-Keyser model (Fig. 10.6) the frontal cyclone develops on a continuous and broad frontal zone (∼400 km) (Fig. 10.6 I). As the cyclone develops, the previously continuous cold front separates or “fractures” from the warm front and advances into the warm sector air, and the temperature gradients of the warm and cold fronts contract inwards (∼100 km) (Fig. 10.6 II). As the cold front moves eastward into the narrowing warm sector, the warm front develops westward resulting in cyclogenesis now occurring in the northerly flow to the west (rear) of the storm within the cold air advancing behind the cold front. The extension of the warm front starts to wrap around in the cold, northerly flow to the rear of the cyclone forming the bent-back front. This bent-back front has the structure of a warm front, and not that of an occluded front as in the classical model. The T-bone term is used as the cold front is orientated perpendicular to the bent-back extension of the warm front (Fig. 10.6 III). Finally, in the warm core frontal seclusion phase (Fig. 10.6 IV), the phase of maximum intensity, the cold front moves further east of the cyclone center (∼500 km). The bent-back warm front and the cold polar air encircling the low, partially or totally enclose a pocket of relatively warm air at its center, thus forming a warm-core seclusion. This seclusion forms within the polar air, and unlike the Norwegian Cyclone model, does not include air originating from the warm sector.
While some observational studies over ocean (e.g. Wakimoto et al., 1992; Neiman and Shapiro, 1993; Neiman et al., 1993; Blier and Wakimoto, 1995) and land (Martin, 1998a,b) have demonstrated the features described by the Shapiro-Keyser model, others have found the Norwegian cyclone model more applicable, or that different elements of both models may be applicable to the same storm (e.g. Mass and Schultz, 1993). Several idealized (e.g. Hoskins, 1976; Hoskins and West, 1979; Polavarapu and Peltier, 1990) and case study simulations (e.g. Anthes et al., 1983; Kuo et al., 1991) have also exhibited evolutions and structures that resemble the Shapiro and Keyser model, while other model simulations, in particular, those conducted over land, or those using surface friction values representative of land, have found the resultant structure more representative of the Norwegian cyclone model (Kuo et al., 1992; Schultz and Mass, 1993). Neiman and Shapiro (1993) emphasize that the frontal evolution described by the Shapiro-Keyser model is not characteristic of all maritime cyclogenesis, and that this model should be considered a companion to, rather than a replacement of, the classical Norwegian cyclone model. Mass and Schultz (1993) speculated that the differences they observed between the storm and the Keyser-Shapiro model may be attributed to the differences in surface properties of land and ocean. Martin (1998a,b) also suggested that the
evolution of the bent back front may be different over land from that over the ocean, although the structure is more robust above the friction layer, implying that there may be greater similarities between the frontal structure of marine and continental cyclones than was previously thought. It thus appears that both the Norwegian cyclone model and the Shapiro-Keyser model should be considered when examining the development of extratropical cyclones.

Recently CloudSat (Stephens et al., 2002) transects through a cold front, warm front and occluded front of three different maritime extratropical cyclones were used to examine the internal structure of frontal clouds and precipitation (Posselt et al., 2008). These transects are shown in Fig. 10.7. The transect through the cold front case (Fig. 10.7a) shows a classic cold frontal structure. Shallow convection is evident in the cold and relatively unstable air to the north of the front, and deep convection can be seen associated with the narrow region of instability at the front’s leading edge. The general cloud distribution is in keeping with the Norwegian cyclone model, however, the CloudSat reflectivity data shows further details of the internal cloud structure not included in the model. For example, multiple precipitating low-level convective showers are evident in the cold air behind (to the north of) the front; the cirrus cloud does not decrease uniformly ahead of the front, but rather shows pockets of high reflectivity, suggesting the presence of multi-cellular convection; and most of the shallow convective clouds in the weakly stratified cold air behind the front appear to be producing precipitation at the surface.

The cross-section through the warm front case (Fig. 10.7b) demonstrates a broad area of cloud and precipitation along and ahead of the warm front itself. The classic sloping distribution of clouds associated with the warm front is evident, and the general cloud distribution is similar to that of the Norwegian cyclone model. As with the cold front, the CloudSat data once again reveal greater details of the warm front cloud and precipitation structures, including that instead of sloping upward with height as in the Norwegian cyclone model, the observed cloud tops remain relatively constant. Large variability in the reflectivity, and hence cloud water content, may be observed. Also, the enhanced region of reflectivity near the base of the cloud with increasing distance from the warm front indicates that large ice particles are settling into the lower regions of the cloud. Finally the extensive region of cloud to the south of the surface warm front produces less precipitation than the clouds directly on top of the warm front, despite similar reflectivity values. This is part of the warm conveyor belt airstream (discussed below), which was not included in the Norwegian cyclone model.

While the cold and warm front transects show characteristics similar to those described in the Norwegian cyclone model, the cross section through the occlusion (Fig. 10.7c) demonstrates a cloud distribution that is significantly different from the classical model, and more in keeping with the Shapiro-Keyser
FIGURE 10.7 CloudSat observed radar reflectivity (dBZ, color shading) overlaid with ECMWF-analyzed equivalent potential temperature (K, solid red lines) for transects through a (a) cold front (0448 UTC 22 November 2006), (b) warm front (1730 UTC 22 November 2006), and (c) the occluded sector (0553 UTC 5 December 2006) of three different mid-latitude cyclones. Note that the ground clutter tends to obscure the radar signal from the cloud below 1 km above the surface. The positions of the front and tropopause are marked with a heavy black line. CloudSat-estimated precipitation rates are depicted at the base of the plot. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.) (Adapted after Posselt et al. (2008))

model. The cloud and precipitation fields are similar to those of the warm front, although the precipitation rates are lower. These observations demonstrate, as stated above, that both the classic and the Shapiro-Keyser models should be considered when examining the structure and life cycles of extratropical cyclones.
Chapter 10 Extratropical Cyclones and High Clouds

10.3.3. Conveyor Belt Concepts

Browning and his collaborators observed similarities in the mesoscale cloud and precipitation features of extratropical storms over the United Kingdom and began to relate these features to commonly observed airstreams or conveyor belts in these storms (Browning and Harrold, 1969; Browning, 1971; Browning et al., 1973; Harrold, 1973). They focused specifically on the warm, ascending airstream that flowed over the warm front, and a drier mid-tropospheric flow that capped the warm, ascending air, thereby generating potential instability as the two airstreams rose over the warm front. They also identified a dry, cool airstream that descended from the downstream anticyclone and flowed under the warm front. Palmén and Newton (1969) had also depicted three airstreams in their study of extratropical cyclones. Schultz (2001) offers an excellent review of the development of these early airstream models. Carlson (1980) extended these airstream ideas to develop a conveyor belt model for extratropical cyclones. His model, shown in Fig. 10.8, consists of three streams of air: (a) the warm conveyor belt, (b) the cold conveyor belt and (c) the dry airstream.

10.3.3.1. The Warm Conveyor Belt

The warm conveyor belt is a stream of relatively warm moist air that originates in the low levels of the southeast quadrant of the storm and flows northward and westward toward the center of the cyclone (in the Northern Hemisphere). As isentropic surfaces slope upward to the north, this air rises as it flows through

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**FIGURE 10.8** The conveyor-belt model of airflow through a northeast US snowstorm. (Adapted from Kocin and Uccellini 1990, Fig. 26, based on the Carlson (1980, Figs 9 and 10) conceptual model [after Schultz (2001)])
the warm sector and above the surface warm front. Interpreted in terms of the quasigeostrophic omega equation [Eq. (10.5)] on constant-pressure surfaces this air, which is ahead of an advancing upper-level trough, is usually associated with positive vorticity advection and warm air advection.

If the air in the warm conveyor belt is convectively unstable, the lifting will destabilize the air and lead to convective clouds and precipitation systems, including rain-bands, squall lines, and severe thunderstorms. These systems are often most intense along and ahead of the cold front where ageostrophic vertical circulation and frictional convergence at the surface cold front produces maximum updraft speeds. If the air in the warm conveyor belt is absolutely stable, then lifting will produce extensive layers of stratiform clouds, including nimbostratus. Often parts of the warm conveyor belt will be convectively unstable, such as in the low levels in the warm sector, while other parts will be absolutely stable, such as in the upper levels of the cool sector north of the surface warm front. Thus, convective clouds and precipitation systems are most likely in the warm sector, while stratiform clouds and precipitation are predominant in the cool sector. Diabatic heating is important in modifying the air in the warm conveyor belt. In the low levels, surface heating destabilizes the air when sufficient solar radiation is available, and fair-weather cumulus clouds may populate this region during the daytime. When either convective or nonconvective precipitation occurs in the warm conveyor belt, latent heat release reduces the adiabatic cooling of the rising air and represents a significant energy source to the cyclone.

The air in the warm conveyor belt eventually reaches the upper troposphere and, according to Carlson (1980), turns anticyclonically toward the northeast, representing the southwesterly flow ahead of the upper-level trough on a constant-pressure surface. More recently it has been argued that, during the early stages of cyclogenesis, if the upper-level flow is an open wave, then the warm conveyor belt turns anticyclonically at jet level, but that later if the upper-level flow is characterized by closed flow, then some of the airflow turns cyclonically around the low center (Bader et al., 1995; Browning and Roberts, 1996; Wernli, 1997). This flow is the trowal airstream discussed in further detail below. Mass and Schultz (1993) also observed that rising trajectories within the warm sector (corresponding roughly to the warm conveyor belt) appear to fan out as they ascend, with some turning cyclonically to the west and others anticyclonically to the east. Such a cyclonic flow regime allows for the contribution of the warm conveyor belt to the comma-shaped cloud mass. By the time the air reaches the upper troposphere, it is usually absolutely stable, so the clouds are extensive sheets of altostratus and cirrostratus. Downstream of the cyclone, the leading edge of the warm conveyor belt is evident by thin cirrus or cirrostratus clouds (Fig. 10.5).

The left (west) edge of the warm conveyor belt originates farthest south and most closely approaches the cold front. Because of its origin, this air is the warmest, moistest, and most unstable air in the cyclone system. As it approaches
the cold front, it also experiences the greatest rate of lifting. Due to the strong baroclinicity in the frontal region, strong wind shear is present as well. The combination of the moist convectively unstable air, rapid lifting, and strong wind shear makes this region favorable for the severe mesoscale convective phenomena discussed in Chapters 8 and 9. Recently, Eckhardt et al. (2004) presented the first climatology of warm conveyor belts. They found from their analysis that the mean specific humidity at the starting points of warm conveyor belts in various regions varies from 7 to 12 g kg$^{-1}$, and that most of this available moisture is precipitated out, leading to an increase of potential temperature of between 15 and 22 K along warm conveyor belt trajectories. Over a time period of three days, a warm conveyor belt trajectory produces on average about 4 (6) times as much precipitation as a global (extratropical) average trajectory starting from 500m above the ground does.

### 10.3.3.2. The Dry Conveyor Belt

Due to the important role of the dry conveyor belt in convective destabilization and cyclogenesis it has received significant attention (e.g. Carr and Millard, 1985; Young et al., 1987; Browning and Golding, 1995; Browning, 1997). The dry conveyor belt originates in the upper troposphere or lower stratosphere west of the upper-level trough and descends into the middle and lower troposphere. As the air descends, it warms and dries. The northern portion of the dry air stream separates from the descending anticyclonic flow, crosses the trough axis, and flows northeastward parallel to the left edge of the warm conveyor belt, where it then ascends over the warm/occluded front (Fig. 10.8). Although this air stream ascends, it is dry and hence normally cloud free and is sometimes referred to as the **dry tongue** or **dry slot**. The boundary between the dry and warm conveyor belts can be extremely sharp, resulting in a surface feature frequently observed over the High Plains of the United States called a **dryline**. The satellite image in Fig. 10.3 shows a dry tongue and the sharp boundary between the dry and warm conveyor belts. The southern portion of the dry conveyor belt continues to descend, warm and dry as it approaches the surface. Strong surface winds associated with the lower portions of this sinking air stream can produce significant blowing dust in the arid regions of the southwestern United States, south of the cyclone.

### 10.3.3.3. The Cold Conveyor Belt

The stream of air called the cold conveyor belt originates to the northeast of the cyclone. This air flows westward (relative to the cyclone) beneath the warm conveyor belt and north of the warm front. Thus the warm front separates the cold and warm conveyor belts. Over flat terrain, the most pronounced lifting of the cold conveyor belt occurs just north of the cyclone center and under the southwesterly jet aloft. According to Carlson (1980), this usually stable air rises, turns anticyclonically, and flows parallel to the upper regions of the warm conveyor belt, although studies prior to 1980 showed the cold conveyor
belt turning cyclonically around the low center, and then remaining in the lower troposphere behind the cold front. A recent study by Schultz (2001) has shown that the cold conveyor belt has both cyclonic and anticyclonic paths, with the anticyclonic path representing a transition between the warm conveyor belt and the cyclonic path of the cold conveyor belt.

As any precipitation produced within the warm conveyor belt falls through the cold conveyor belt, the temperature and the humidity of the cold conveyor belt can significantly influence the type and amount of precipitation reaching the surface. For example, if the cold conveyor belt is relatively dry, evaporation and/or sublimation of hydrometeors will reduce the amount of precipitation reaching the surface, whereas a nearly saturated cold conveyor belt could result in substantial surface precipitation. Also, if the temperatures of the cold conveyor belt are below freezing, ice pellets or freezing rain may be expected at the surface. Saturation is enhanced by evaporation of precipitation falling from the warm conveyor belt aloft, and hence thick fog and stratus clouds may occur in the cold conveyor belt, north of the surface warm front. Along the eastern slopes of the Rocky Mountains in the United States, the cold conveyor belt experiences significant lifting as it follows the rising terrain, and these “upslope” conditions are responsible for almost all of the nonconvective precipitation in this region (see Chapter 11).

Carlson (1980) concluded that the air in the comma-shaped cloud heads of extratropical cyclones originates in the anticyclonic path of the cold conveyor belt, and Liu (1997) showed that as much as 70% of the moisture transported within extratropical cyclones could be linked to the cold conveyor belt. Others, however, have emphasized the role of the warm conveyor belt in generating heavy precipitation (e.g. Martin, 1999; Schultz, 2001). Schultz (2001) argues that while shallow clouds could be produced within the relatively stable air of both the anticyclonic or cyclonic branches of the cold conveyor belt, heavy precipitation associated with strong vertical uplift and deep clouds is not likely to be produced within the stable environment of the cold conveyor belt. The latter is more likely to be generated at mid-levels by the cyclonic path of the warm conveyor belt. The cool cyclonically flowing low-level air to the rear of the extratropical cyclone is however frequently moist enough that the frictionally-induced rising motion may produce an extensive layer of stratocumulus or fair-weather cumulus clouds at the top of the PBL.

The conveyor-belt model has been criticized in the past for a number of reasons, a detailed overview of which is provided by Schultz (2001). Some studies have found no evidence of a well-defined cold conveyor belt (e.g. Reed et al., 1994; Browning et al., 1995), sharp boundaries between the warm and cold conveyor belts (e.g. Kuo et al., 1992; Browning and Roberts, 1994), or any anticyclonic turning of the cold conveyor belt (Mass and Schultz, 1993). Concerns have been expressed in the literature as to whether the conveyor belts represent streamlines, trajectories, streak lines or some combination of these.
It has been suggested that the airstreams cannot be represented as steady-state, flat belt-like structures or three-dimensional tubes with well-defined boundaries due to their complex geometries, and should rather be thought of as flexible tubes that evolve over time (Kuo et al., 1992; Reed et al., 1994; Wernli, 1997). It has also been argued that the three-belt model is an oversimplification of the airflow, with more than three air streams being needed to show the various parcel trajectories (Mass and Schultz, 1993; Reed et al., 1994; Bader et al., 1995). In spite of these criticisms, the conveyor-belt model does provide some useful guidance as to the basic airflow through extratropical cyclones and continues to be utilized.

10.3.4. Frontal Occlusion and the Trowal Airstream

The concept of occluded fronts, in which a cold front overtakes a warm front thus lifting the warm sector air above the frontal intersection, was first introduced within the Norwegian cyclone model. In an ideal warm (cold) occlusion, the air preceding the warm front is colder (warmer) than the air behind the cold front so that the cold (warm) front rides aloft over the warm (cold) front. More recently, Schultz and Mass (1993) defined the occlusion process as one in which the low center becomes progressively separated from the warm sector of the cyclone resulting in a pressure drop, and an associated tongue of intermediate temperature air that extends from the low center to the warm sector. Debate continues to exist in the literature regarding the occlusion process. Some have suggested that the important feature of an occlusion is the trough of warm air that is lifted aloft ahead of the cold front rather than the surface occluded front (e.g. Crocker et al., 1947; Godson, 1951; Penner, 1955; Galloway, 1958, 1960). Others have shown that an occluded structure can occur without the classical occlusion process occurring (e.g. Reed, 1979; Locatelli et al., 1989), or that there is little evidence that supports the occlusion process as described in the Norwegian cyclone model (e.g. Shapiro and Keyser, 1990). Still others have found that while the surface-based cold front does catch up to the warm front it does not ride over the warm front, but that the upper-level frontal zone appears to ascend over the surface-based warm front (Schultz and Mass, 1993). Posselt and Martin (2004) described a case in which the occlusion first occurred at mid-tropospheric levels and then at the surface. Finally, Schultz and Mass (1993) suggest that warm-type occlusions, rather than the cold-type occlusions, are the normal structural form.

As just stated, one of the main results of the warm occlusion is the production of a wedge of warm air aloft, which is displaced poleward of the surface warm and occluded fronts, as is shown schematically in Fig. 10.9a. This trough of warm air, referred to as the trowal (trough of warm air aloft), was observed to correspond more closely to the cloud and precipitation characteristics of the occluded cyclone than to the surface occluded front (Crocker et al., 1947; Godson, 1951; Penner, 1955; Galloway, 1958, 1960).
FIGURE 10.9  (a) Schematic illustration of the trowal conceptual model. The lightly shaded surface represents the warm edge of the cold frontal baroclinic zone. The darker shaded surface represents the warm edge of the warm frontal baroclinic zone. The thick dashed line (marked “TROWAL”) represents the three-dimensional sloping intersection between the cold and warm frontal zones characteristic of warm occlusions. Schematic precipitation band is indicated as are the positions of the surface warm, cold and occluded fronts (after Martin, 1999).  (b) Elevated view from the north of the 309 K $\theta_e$ surface from an 18-h forecast of the UW-NMS valid at 1800 UTC 19 January 1995. The dashed line indicates the trowal position, which slopes upward from near the surface to 8 km. Position of the surface cold front is also indicated. (c) Equivalent potential temperature at 2 km from an 18-h forecast of the UW-NMS valid at 1800 UTC 19 January 1995. Solid lines are $\theta_e$ labeled in K and contoured every 4 K. Thick dashed line is the axis of maximum $\theta_e$ characteristic of a warm occluded cyclone and used to approximately locate the trowal. (Adapted after Martin (1999))
While the trowal was initially indicated on a surface map showing the intersection of the cold and warm front aloft (Crocker et al., 1947; Godson, 1951; Penner, 1955), the trowal concept has evolved over time to represent the warm anomaly aloft (Martin, 1998a,b, 1999; Market and Cissell, 2002; Posselt and Martin, 2004; Moore et al., 2005). The trowal may be identified either as a ridge of high (equivalent) potential temperature on a horizontal cross-section, or as a three-dimensional sloping canyon on an isosurface of (equivalent) potential temperature, as shown in Fig. 10.9 (b, c). Martin (1998a) defined the trowal as marking the “3-D sloping intersection of the upper cold-frontal portion of the warm occlusion with the warm-frontal zone”. The upward-sloping tongue of warm air wraps cyclonically around the cyclone.

In satellite imagery of maturing continental cyclones it is evident that the northwestern boundary of the dry slot typically coincides with the southeastward boundary of the cloud pattern produced in association with the trowal airstream (Grim et al., 2007). A number of numerical simulations of occluded cyclones (Schultz and Mass, 1993; Reed et al., 1994; Martin, 1998a,b) have exhibited the thermal characteristics of the occluded structure just described. They have also shown a cloud- and precipitation-producing airstream that originates in the warm sector boundary layer and that ascends cyclonically in the occluded quadrant of the cyclone, which Martin (1998b) called the trowal airstream. Martin (1999) found that convergence of the along-isentrope component of the Q-vector simultaneously creates the thermal ridge and provides most of the Q-G forcing for upward motion within the occluded quadrant and of the trowal airstream. The trowal airstream is associated with the wrap-around cloud and precipitation that is commonly observed with occluded extratropical cyclones, and has been linked to heavy snowfall produced by intense snowbands, as will be seen below.

10.3.5. Cloud and Precipitation Characteristics

10.3.5.1. CSI and Precipitation Bands

As will be seen in the next section, clouds and precipitation associated with extratropical cyclones tend to be organized into bands. The roles of frontogenesis and CSI in producing mesoscale precipitation bands, and the fact that these processes may operate together to generate such bands have received significant attention in the literature (e.g. Bennetts and Hoskins, 1979; Emanuel, 1979, 1983a,b, 1985; Sanders and Bosart, 1985; Sanders, 1986; Xu, 1989; Shields et al., 1991; Nicosia and Grumm, 1999). Emanuel (1979, 1983a,b, 1985) showed that rising motion in association with frontogenesis was contracted and enhanced when CSI or even weak symmetric stability were present, thus suggesting a relationship between frontogenesis and symmetric stability in producing precipitation bands. Xu (1989) demonstrated that frontogenesis occurring in the presence of CSI can, in theory, produce long-lived precipitation bands. Moore and Lambert (1993) developed a two-dimensional form of
equivalent potential vorticity (EPV), later extended to a more general, three-dimensional form by McCann (1995), that could be used to diagnose regions of CSI. They found that CSI could be diagnosed in a region of negative EPV, a region which tends to be saturated and characterized by strong vertical wind shear and weak convective stability.

The relationship between frontogenesis, CSI and EPV has been discussed in detail by others such as Bluestein (1993) and Nicosia and Grumm (1999), and is summarized here. As just stated, CSI is present when EPV is negative; CI may also be present in conjunction with CSI when EPV is negative. For a layer characterized by negative EPV, CSI and/or CI can only be released if the layer becomes saturated in association with rising motion. Also, when CSI and CI are both present, it is expected that CI-associated upright convection will dominate over the CSI-associated slantwise convection. It can be shown for frictionless, adiabatic flow that EPV will be reduced in a region where the moisture gradient lies in the same direction as the thermal wind vector. Such a scenario regularly occurs for extratropical cyclones located within a baroclinic zone with the cold air to the north. In this scenario, moist air is found to the east, while the drier air tends to be located to the west. When frontogenesis occurs in a region of negative EPV and CSI, stronger vertical motion develops and becomes constricted to a smaller scale (Emanuel, 1979, 1983a,b, 1985). Stronger vertical motion leads to stronger convergence in the lower levels which enhances the temperature gradient and hence frontogenesis. This in turn increases the temperature gradient, which further reduces the EPV. The continual decrease in EPV results in the development of CSI, and CI to a lesser extent, in regions of frontogenesis. As the rising air reaches saturation, CSI and CI are released, thereby enhancing the vertical motion within the region of frontogenesis and subsequent band development. In this way a positive feedback is established.

As Nicosia and Grumm (1999) pointed out, the fact that frontogenesis occurs in or near a region of CSI is not just a coincidence. In regions of frontogenesis, the potential temperature gradient increases, which requires an increase in the geostrophic wind shear through thermal wind arguments. Such geostrophic wind shear produces differential moisture advection and a subsequent increase in the slope of the \( \theta_e \) isentropes. The weakening of the convective stability and the strengthening of the vertical wind shear results in negative EPV and CSI. The reduction in EPV can take place at quite a distance from the surface cyclonic circulation (Moore et al., 2005). It is the unique juxtaposition of the three conveyor belts within an extratropical cyclone that results in a region of negative EPV and associated CSI within a region of midlevel frontogenesis (Nicosia and Grumm, 1999). EPV is significantly reduced on the warm side of the midlevel frontogenesis due to the passage of the dry conveyor belt over the cold conveyor belt. The continual reduction in the EPV produces a deep layer of negative EPV located within a region of warm advection and uplift associated with the warm conveyor belt.
FIGURE 10.10  Visible satellite photograph of occluding extratropical cyclone over North Pacific. Note cellular cumulus convection south of cyclone center and small cloud system southeast of center behind cold-frontal band of clouds. This system is an incipient mesoscale cyclone. Time is 1845 GMT, 27 February 1980.

10.3.5.2. Rainbands

The mesoscale organization of clouds and precipitation within extratropical cyclones has been extensively studied using both in situ and remotely-sense observations, as well as numerical simulations (e.g. Browning and Harrold, 1969; Browning, 1971; Browning et al., 1973; Houze et al., 1976; Hobbs and Locatelli, 1978; Herzegh and Hobbs, 1980; Matejka et al., 1980; Houze and Hobbs, 1982; Hobbs and Persson, 1982; Houze, 1993; Braun et al., 1997; Jorgensen et al., 2003). A summary of the early history of the research leading to our knowledge of the mesoscale precipitation structure of extratropical storms is provided by Atkinson (1981). These studies have shown that clouds and precipitation tend to be organized in band-like structures referred to as rainbands (discussed here) and/or snowbands (discussed in the next section). An example of such bands may be seen in Fig. 10.10, which shows an occluding extratropical cyclone over the North Pacific. As the upper-level trough deepens and becomes a closed cyclone, the low- and middle-level clouds are wrapped around the cyclone center. A region of organized, cellular cumulus convection is present in the unstable PBL south of the cyclone center. Bands of convective clouds occur along the cold front and in the convectively unstable warm sector ahead of the front. Rainbands are typically orientated parallel to one of the fronts, contain smaller regions of more intense precipitation, are found throughout extratropical
cyclones, and are classified as wide (50-75 km) or narrow (5-25 km) according to their width.

Six different types of rainbands typically associated with Pacific storms have been identified in a classification scheme developed by Hobbs (1978), Matejka et al. (1980) and Houze and Hobbs (1982). The scheme, which we will refer to as the HHM scheme, was based on the location and the morphology of the bands, the location of which is shown in Fig. 10.11. The six types of rainband include the following:

1. **Warm frontal rainbands:** Warm frontal bands, depicted in cross-sectional form in Fig. 10.12, are oriented parallel to the surface warm front where a deep layer of ascending warm, moist air exists. They are typically of the order of 50 kilometers wide and several hundred kilometers long. Ice particles may form above the warm front, often within shallow convective cells, and may
2. **Warm sector rainbands:** Warm sector bands (~50 km wide) are located ahead of and parallel to the cold front, and often resemble squall lines. Younger convective bands in this region tend to occur ahead of older, less convective bands, and contain both liquid water and ice compared with their nearly glaciated, older counterparts.

3. **Narrow cold frontal rainbands:** A narrow (~5 km) cold frontal rainband (NCFR) is typically observed at the leading edge of the surface cold front and is probably caused by intense frictionally induced convergence at the front (Keyser and Anthes, 1982). NCFRs are characterized by vigorous, narrow convective updrafts and narrow downdrafts, are often associated with heavy convective rainfall (~10-50 mm hr\(^{-1}\)) and may produce severe weather. In spite of the strength of the updrafts associated with the NCFRs (Carbone (1982) observed updrafts up to 18 m s\(^{-1}\)), which tend to be nearly vertical, the rainbands typically only extend to heights of between 3.5 and 5 km. They appear to behave like density currents in which the vertical circulation is characterized by an approximate balance between the baroclinically generated vorticity of the cold front and the ambient vertical wind shear (Rotunno et al., 1988; Parsons, 1992; Jorgensen et al., 2003). Sometimes the NCFR
occurs near the leading edge of a larger area of stratiform rainfall induced by post-frontal, slantwise convection, although it appears that they are more typically embedded within the large stratiform cloud region; occasionally they may be neither (Koch and Kocin, 1991). It should be noted here that a front having the local character of a gravity current is called a front with a gravity current-like structure (Smith and Reeder, 1988). We will use the term density current to refer to the density difference across a cold front, and the term gravity current-like structure to refer to a cold front that locally has the characteristics of a gravity current. Gravity currents can propagate ahead of cold fronts, where they may be associated with warm sector rainbands.

4. **Wide cold frontal rainbands:** Wide (~50-75 km) cold frontal rainbands (WCFRs, 50 km in width) are oriented parallel to the cold front and either straddle the front or occur behind the front. In occlusions, they can be associated with the cold front aloft. WCFRs tend to produce regions of enhanced stratiform precipitation. The upward motion of these bands is characterized by mean ascent aloft, although they can also include shallow convective cells that contribute to precipitation through seeder-feeder processes (see below). The WCFR generally forms behind the NCFR but tends to move faster than the NCFR and thus may be co-located with the NCFR or positioned ahead of the low-level cold front. WCFRs may also occur without NCFRs (Evans et al., 2005; Bond et al., 2005). Parsons and Hobbs (1983) describe several cases in which a WCFR moved across a surface cold front. This usually led to a decrease in the frontal convergence and a subsequent reduction in the organization of the NCFR, even occasionally the dissipation of the band. Finally, sub-bands have also been observed within WCFRs (Evans et al., 2005).

5. **Prefrontal cold-surge bands:** Prefrontal cold-surge bands are associated with surges of cold air above the warm front and ahead of the cold front in occluded cyclones.

6. **Postfrontal rainbands:** Postfrontal bands are lines of convective clouds that form in the unstable cold air behind the cold front.

The relatively intense precipitation associated with wide rainbands appears to be generated by a seeder-feeder process, in which the seeding of supercooled water by ice crystals falling from above enhances precipitation (Fig. 10.12). The region where the ice crystals are produced is called the seeder zone. The remaining precipitation is produced in the zone below, termed the feeder zone because of the strong advective supply of moisture by the cyclone-scale circulation. The seeder-feeder process may also enhance the upward motion and precipitation of the other types of bands depicted in Fig. 10.11. The seeder-feeder process has been discussed by Browning and Harrold (1969), Herzegh and Hobbs (1980), and Houze et al. (1981) and modeled by Rutledge and Hobbs (1983), and is described more fully in Chapter 11 with respect to its role in orographic precipitation.
A schematic model developed by Hobbs et al. (1980) depicts the band structure of a typical cold front (Fig. 10.13). In this model, the heaviest precipitation is located in mesoscale rainbands oriented parallel to the cold front. These include a warm-sector rainband consisting of a series of mesoscale convective sub-bands, a NCFR, and four WCFRs. Also illustrated are generating zones of ice crystals and the feeder zones which supply moist, cloudy air from which seeder crystals grow by vapor deposition, aggregation, and riming of cloud droplets. Hobbs et al. (1980) estimated that about 20% of the precipitation in the wide cold-frontal rainbands originates in the seeder zones and ~80% originates in the feeder zones.

As reported by Ryan (1996), there are many studies confirming the HHM rainband classification scheme for numerous parts of the world including the United States, the United Kingdom, Russia, Finland, China, Israel and Australia (see Ryan, 1996 for related references), and rainband structures have been observed in Spain, Morocco, Japan and South Africa. The survey conducted by Ryan, although limited, does indicate that the scheme appears valid for a wide range of environmental conditions. However, while the scheme is relatively well established, the mechanisms responsible for the variety of bands in extratropical cyclones are still under debate. This is partly due to the limitations of theoretical models, the variation in the mechanisms from storm to storm and even within the same storm, the range of scales (synoptic, mesoscale, convective, microscale)
involved that make observing and numerically simulating all the processes involved difficult, and the fact that a combination of mechanisms may be operating at once.

A number of studies have focused on the role of different instability mechanisms in generating rainbands, particularly within the warm sector. For example, in more severe cyclonic storms, the warm sector can be conditionally unstable, leading to the formation of prefrontal squall lines, where SI (Ogura et al., 1982) may be responsible for triggering the linear structure of the squall lines. Bennetts and Ryder (1984) compared the predictions of CSI and symmetric wave-CISK (Emanuel, 1982) to the observed banded structure in the warm sector of an extratropical cyclone. The primary energy source for SI is the kinetic energy of the basic flow, while CSI receives additional energy from latent heat release. Symmetric wave-CISK in contrast, derives its energy principally from latent heat release, while additional energy is supplied by the mean flow (e.g. vertical wind shear). Both theories predicted banded rolls, but the CSI theory predicted that the rolls move with the mean wind while the wave-CISK theory predicted that the rolls exhibit a propagation velocity relative to the mean flow of 6 m s$^{-1}$. The observations suggest that CSI theory was more applicable to this case than the wave-CISK theory. As noted in Chapter 8, however, a number of variations in the parameterizations used in wave-CISK theory can account for the faster propagation of Emanuel’s model. Raymond’s (1983) advective wave-CISK model, for example, predicts a propagation velocity comparable to the mean wind speed. Thus, in the warm sector of many extratropical cyclones, both CSI and wave-CISK may occur, the differences being so slight that variations in the details of the parameterizations used in the models are greater than the fundamental differences between the two theories.

In another extratropical cyclone, Parsons and Hobbs (1983) found that CSI appeared to be the stronger candidate for explaining the observed warm-sector rainbands, and also noted that the CSI theory was consistent with the observed structure of the wide cold-frontal rainbands in the same case. Lemaitre et al. (1989) noticed the role of CSI, as well as other mechanisms, in their analysis of a WCFR over southwestern France. They found that the band was associated with synoptic-scale ascent of the WCB, slantwise convective along the tilted frontal zone and upright convection in the first 10 km at the leading edge of the system. Frontogenetic forcing and CSI appeared to be important to band organization on the mesoscale, while the gravity current-like behavior of the low level cold air produced new cells ahead of the system. Schultz and Knox (2007) examined the development of several mesoscale rainbands that formed over Montana and the Dakotas during July 2005. The bands were located to the north or poleward of a region of frontogenesis. The formation and organization of the rainbands in this case appears to have been dependent on the release of dry symmetric instability with possible contributions from inertial instability within a zone of frontogenesis. However, once the bands began to grow and develop they were able to tap conditionally unstable air.
In Carbone’s (1982) Doppler radar study of the kinematic and thermodynamic structure of a NCFR over California, he found that SI, as well as vertical shear, which occurred in a direction nearly parallel to the band, were likely important factors in the development of this band. This band produced a variety of severe mesoscale weather conditions, including heavy precipitation, strong winds, electrical activity, and tornadoes. The Doppler radar observations showed a nearly two-dimensional updraft of magnitude 15-20 m s$^{-1}$ that was associated with a gravity current that propagated ahead of the cold front toward a strong prefrontal low-level jet. Diabatic cooling associated with the melting of ice was important in maintaining the density contrast across the gravity current. Hobbs and Persson (1982) and Parsons et al. (1987) also proposed that the dynamics of NCFRs are similar to that of a gravity current and that diabatic heating associated with the microphysics of the rainband can influence the density across the gravity current. Condensation in the air rising immediately ahead of the line convection warms the air through the release of latent heat, thereby offsetting the adiabatic cooling associated with lifting. This heating, together with the cooling due to melting and evaporation of precipitation behind the line convection, work in unison to enhance the density contrast across the gravity current. Further warming ahead of the gravity current can occur due to warm air advection by the pre-frontal, low-level jet.

Moncrieff (1989) showed that NCFR dynamics can be represented by steady-state models, and that once a gravity current balance has been generated by precipitation processes, a squall line can be maintained in the presence of negligible potential instability. Shapiro et al. (1985) suggested that non-precipitating fronts may also develop gravity-current-like features, however, the observed circulations were not typical of those that occur in association with classical atmospheric gravity currents. A similar concern was raised by Smith and Reeder (1988) regarding the precipitating studies of Browning and Harrold (1970), Testud et al. (1980) and Carbone (1982). In the idealized study of Knight and Hobbs (1988) a NCFR was generated by frictional convergence in the PBL that was free of precipitation, and that showed no similarities to a gravity current.

Others have also examined the diabatic effects of condensation and evaporation on rainband formation. Hsie et al. (1984) used an explicit model for clouds and precipitation to study the diabatic effects of condensation and evaporation on an idealized model of a cold front. Without moisture, frontogenesis occurred in the two-dimensional model as a result of geostrophic shearing deformation. Upward motion occurred just ahead of the surface cold front (SCF) (Fig. 10.14) as a result of deformational and frictional processes. A northerly jet developed at the top of the PBL behind the SCF, while a low-level southerly jet developed in the warm sector ahead of the SCF.

The moist simulation contained some distinctive features not present in the dry simulation. The vertical velocity showed a banded structure in the warm sector (Fig. 10.15). The first band was associated with the frictional convergence
around the pressure minimum at the SCF, and was stronger than the updraft in the dry case (Fig. 10.14). The second and third bands were associated with moist convection; these bands formed in a region of CSI. The horizontal wavelength of 200-300 km of the convective bands is similar to the shorter wavelength mode predicted by Emanuel (1982) for CSI. Initially, the formation of the cloud was from the large-scale upgliding component of the frontal circulation. Later, this large-scale motion broke down into the banded structure. The narrow band that formed in the low levels close to the surface cold front (band 1 in Fig. 10.15) is similar to the NCFR and moved with the same speed as the SCF (Hobbs et al., 1980). The wider band straddling the cold front at higher levels (centered at $x = 180$ km and $z = 4.5$ km in Fig. 10.15) is similar to the WCFR. Two other bands resembled observed warm sector bands. Both the warm-sector and WCFRs moved at a speed faster than the SCF. The behavior of the bands in the moist simulation agrees qualitatively with the observations of Hobbs et al. (1980).

In addition to producing the rainbands and a stronger low-level jet, the effects of latent heat on the cold frontal circulations are summarized as follows:

1. Latent heating produces a stronger horizontal potential temperature gradient across the front, especially in the middle and upper levels. The gradient in the low levels is relatively unaffected.
2. Convection results in a stronger static stability across the front and a weaker static stability in the convective region.
FIGURE 10.15  As in Fig. 10.14, but for moist simulation. Cloud boundary is indicated by a thick dash-dot line; position of surface cold front (SCF) is indicated by vertical arrow at about 160 km. Shaded areas denote regions of negative equivalent potential vorticity. (From Hsie et al. (1984))

3. Both the upper- and lower-level jets are stronger when latent heating is present. The horizontal wind shear (or relative vorticity) is stronger in the moist simulation, especially in the low levels.

4. Convection intensifies the ageostrophic circulation around the frontal zone. The circulations in the warm sector appear to be dominated by convection.

5. Moist convection increases the speed of the SCF and reduces the slope of the front.

Rutledge (1989) found from his 3-D cloud simulation of a NCFR that diabatic processes were important in the maintenance of the cross-frontal temperature gradient, and that a positive feedback exists between cloud microphysical processes of the rainband and the maintenance of the density current. The diabatic processes of evaporation, melting and condensation associated with the rainband enhance the density current as described above. The enhanced density current, in turn, drives a stronger vertical circulation, thereby maintaining the rainband. Rutledge suggested that cooling associated with the melting of ice may also contribute to the preservation of the density current, as the stable layer generated by melting could limit the dissipative effects of shear-instabilities and wave-breaking behind the head of the gravity current.

Barth and Parsons (1996) investigated the microphysical processes within a NCFR through the use of a 2-D nonhydrostatic cloud model with ice microphysics. Their results showed that intense but shallow updrafts along the NCFR produced significant amounts of cloud water which led to rain and
graupel, and, to a lesser extent, snow, within the region of heavy precipitation of the NCFR. The ice phase was needed in order to represent the stratiform precipitation produced by this band. Like Rutledge (1989), they also found positive feedbacks between the microphysics and the dynamics of the NCFR, with melting and sublimation enhancing the cooling within the cold air mass, thus intensifying the circulations that support rainband development. However, their simulations indicated that the prime role of these ice phase processes was to enhance cooling and hence baroclinicity across the front, as opposed to decreasing the mixing across the frontal air mass through the melting-induced stabilization.

A NCFR associated with a cold front with a gravity current-like structure was examined by Koch and Kocin (1991) and Chen et al. (1997) using observational and numerical model output, respectively. This case was characterized by frontal merging and strong frontal contraction that generated severe line convection, and that had a gravity current-like structure along the leading edge of the surface cold front. The NCFR developed in an environment devoid of potential instability, thus requiring significant forced lifting. Koch and Kocin suggested that, while frontal merger processes and hydraulic jumps generated by the flow over the Appalachian mountains were potential mechanisms for forcing this uplift, they had insufficient data to address this. Their analysis did however suggest that the gravity current-like structure of the cold front that developed following the formation of the NCFR may have arisen from the evaporation and melting of hydrometeors. Strong lifting of the air may thus have been possible due to a near balance between the solenoidal circulation associated with the negative vorticity of the gravity current and the circulation associated with the positive vorticity of the vertical shear of the low-level jet (Rotunno et al., 1988).

Chen et al. (1997) simulated this case in order to determine whether the gravity-current-like structure was generated by precipitation melting and evaporation. Their simulations revealed that neglecting the moist processes in the model did not have a significant effect on the gravity-current-like structure of the front, and that planetary boundary layer frictional processes appear to play a much greater role. These frictional processes generate cross-frontal low-level wind shear, which produces low-level convergence and steepens the isentropes along the leading edge of the front. Thus the isolines of $\theta_e$ may become steeper than the absolute momentum surfaces, resulting in negative PV in the planetary boundary layer. A negative PV anomaly would inhibit the geostrophic adjustment process, thereby maintaining a front with gravity-current-like structure. Therefore, the observed, intense NCFR was primarily related to frictionally-induced PBL processes.

Others have investigated the role that PBL processes can play in the formation of rainbands. Keyser and Anthes (1982) found that PBL processes can be important in altering ageostrophic circulations with fronts, thus producing a narrow intense updraft at the warm edge of the front through frictionally induced
convergence. Reeder (1986) noted that differential heating in the PBL resulted in the development of an intense prefrontal updraft, and that a gravity-current-like structure developed at the leading edge of the front. Knight and Hobbs (1988) observed from their modeling study, in which precipitation processes were included, that the formation of NCFRs was aided by frictional convergence in the PBL. PBL processes were also found to be important in the generation of NCFRs by Benard et al. (1992) and Koch et al. (1995).

Four types of rainbands, two narrow and two wide, were identified by Benard et al. (1992) in their 2-D nonhydrostatic simulation of moist frontogenesis in an idealized baroclinic wave. A NCFR, positioned at the surface cold front, produced heavy precipitation and consisted predominantly of a line of shallow convection that was triggered by frictionally induced instability. Narrow free-atmosphere rainbands driven by a series of updrafts and downdrafts were located above the convective line of the NCFR. Gravity wave theory, rather than CSI or conditional convective instability, appears to explain the characteristics of these bands. WCFRs were found to repeat periodically in the frontal zone, with a lifetime limited to six to nine hours, and appeared to be associated with slantwise convection. A single warm-sector rainband was identified at 300 to 400 kilometers ahead of the surface cold front and produced widespread precipitation.

Finally, Szeto et al. (1999) described the role of the surface in their simulations of a severe ice storm associated with the passage of a warm front over the east coast of Canada. Their goal was to understand the cloud and mesoscale processes that affected the development of freezing rain in this system. One of the mechanisms for the formation of freezing rain is the presence of an above-freezing inversion layer (AFIL) at low levels with a layer below the AFIL of subzero temperatures. Such low-level inversions are often associated with surface fronts, however, the dynamics driving the AFIL are not well understood. This is further complicated by the fact that cooling produced by melting within the warm inversion layer, and warming associated with refreezing in the lower levels, can alter this vertical temperature structure. Analysis of their model results showed that as the surface warm front approached Newfoundland, the change in surface characteristics from ocean to land disturbed the quasi-thermal wind balance, the result of which was accelerated warm frontogenesis through the intensification of the ageostrophic cross-frontal circulation. This, in turn, produced an extensive AFIL, which may be associated with ice pellets or freezing rain depending on the depth of the subzero layer. Potential instability and CSI appeared to be the mechanisms that produced the banded precipitation features within the model.

10.3.5.3. Snowbands

Winter cyclones, whether explosively deepening (e.g. Schneider, 1990; Powers and Reed, 1993; Marwitz and Toth, 1993; Mass and Schultz, 1993; Pokrandt et al., 1996), or less intense (e.g. Hakim and Uccellini, 1992; Shea and
Przybylinski, 1993), are often accompanied by a combination of strong winds, freezing rain, subfreezing temperatures, heavy snow and blizzard conditions. Modest intensity extratropical cyclones make up the majority of all cyclone events (Roebber, 1984). These cyclones can produce significant amounts of precipitation through a variety of forcing mechanisms that influence the precipitation processes but that may not all simultaneously contribute to the cyclone development (Martin, 1998a,b). This situation occurs frequently in the central United States where the synoptic-scale forcing does not generate a strong surface disturbance, and yet large amounts of low-level moisture and a weakly stratified lower troposphere can result in a significant response to frontal forcing. Such a case occurred over the central United States in January 1995 when moist, relatively warm low-air from the Gulf of Mexico, and warm frontogenesis processes, produced heavy snow in association with snowbands that developed parallel to the warm front (Martin, 1998a,b).

Snowbands are intense, narrow (5-40 km) bands of heavy snow that are typically located within larger (100-500 km wide) regions of light to moderate snow. Numerous studies have been conducted on the factors affecting the development of snowbands (e.g. Sanders, 1986; Martin, 1998a,b; Nicosia and Grumm, 1999; Schultz and Schumacher, 1999; Jurewicz and Evans, 2004). A number of different processes have been identified that may enhance snowband formation including jet streak interactions, frontogenesis in the presence of weak moist symmetric stability or convective instability, and local orographic forcing. Research by Novak et al. (2004) and Martin (1998a,b) indicated that snowbands producing heavy snowfall generally occur in regions where frontogenesis and instability are co-located, typically to the northwest of surface cyclones in large-scale flow that is highly amplified. However, several more recent studies demonstrate that less intense, but still significant snowbands can occur in other locations relative to a surface cyclone, and in other types of large-scale environments such as to the northeast of the surface cyclone (e.g. Banacos, 2003; Schumacher, 2003) or in situations of no surface cyclone (e.g. Skerrett et al., 2002).

An explosively developing extratropical cyclone occurred over the United States during December 1987. This record-breaking snowstorm, which moved from New Mexico to the Great Lakes region between 13 and 16 December producing heavy snowfall along its path, has been the subject of numerous studies (e.g. Schneider, 1990; Powers and Reed, 1993; Pokrändt et al., 1996). Over a foot of snow over Oklahoma was produced by a northeast-southwest orientated warm-frontal snowband. Marwitz and Toth (1993) studied the mechanisms producing this snowband and found that frontogenesis was occurring in the warm frontal region, and that a direct circulation occurred around the warm front. Above the warm front, ageostrophic winds forced the conditionally unstable air to rise, thereby releasing its instability. Thus both frontogenetic forcing and convective buoyancy were responsible for forcing the snowband.
Martin (1998b) examined a long, narrow snowband (∼1100 km) that developed in association with a moderate extratropical cyclone on 19 January 1995. The snowband produced thunder, lightning, heavy winds and record-breaking snowfall. Similarly to Marwitz and Toth (1993), Martin concluded that the snowband was forced by the direct vertical circulation associated with warm frontogenesis. The vertical circulation was supplied with high-θ_e air from the Gulf of Mexico, and the release of convective instability within the ascending branch of the vertical circulation resulted in the convective characteristics of the band. Lifting of warm, moist air within the trowal also contributed to the heavy snowfall, with frontogenesis along the warm-front portion of the occluded structure being the mechanism by which the air was lifted into and through the trowal. While the precipitation bands were orientated parallel to the warm front, thus suggesting that the release of CSI may have played a role in the band formation, saturated regions of CSI could not be found in the region of the warm front within the model, although the model resolution was relatively coarse.

As discussed above, CSI, and at times CI, are present when EPV is negative in a baroclinic atmosphere. When low EPV occurs in a region of frontogenesis, stronger vertical motion develops and becomes constricted to a smaller scale (Emanuel, 1985). Stronger vertical motion leads to stronger convergence in the lower levels which enhances frontogenesis and the temperature gradient. The increased temperature gradient is associated with further reductions in the EPV. The continual decrease in EPV results in the development of CSI, and CI to a lesser extent, in regions of frontogenesis. As the rising air reaches saturation, CSI and CI are released, thereby enhancing the vertical motion and the associated snowband development. The enhanced vertical motion within the frontogenetic region enhances low-level convergence and a positive feedback is established.

Nicosia and Grumm (1999) made use of both model output and radar data to examine intense mesoscale band formation in three northeastern snow storms associated with extratropical cyclones. The heaviest total snowfall rates in each of these storms was co-located with the location of these mesoscale snowbands. Rates reached as high as 6 in h\(^{-1}\). Their analysis showed that the mesoscale snowbands formed in a region to the north of the developing cyclone at midlevels, which was characterized by intense midlevel frontogenesis and a deep layer of negative EPV. The location of the dry conveyor belt over the moisture-rich cold conveyor belt to the north of the low-level cyclone resulted in the significant reduction of the EPV. As air parcels ascended north over the warm front CSI, and CI to a lesser extent were released upon saturation, resulting in enhanced vertical motion and mesoscale band formation. As both CSI and CI were present, it is likely that upright convection associated with the release of CI dominated over the slantwise convection associated with the release of CSI, thereby generating heavier snowfall rates. Therefore the mesoscale snowbands in these cases formed in association with intense midlevel frontogenesis and a deep layer of negative EPV.
Jurewicz and Evans (2004) compared two banded snowstorms that occurred over the northern mid-Atlantic region during January 2002 under very different synoptic-scale forcing. They found that in both cases the snowbands were co-located with mid-tropospheric frontogenesis and reduced stability. In the first case, a layer of PSI occurred just above a deep sloping frontogenetic zone with near-saturated conditions, while in the second case, a layer of PI occurred just above a shallow sloping frontogenetic zone in association with decreasing humidity with height. Jurewicz and Evans concluded that the difference in the snowband characteristics arose from differences in the synoptic-scale forcing, the frontogenetical forcing, the amount of moisture available, the degree of instability and the thermal profiles.

Moore et al. (2005) examined the processes associated with a long (1000 km), narrow snowband that produced heavy snowfall from the Texas panhandle to northwest Missouri on 4-5 December 1999. Thundersnow was also reported. The snowband was located along a northeast-southwest frontal boundary several hundred kilometers to the northwest of a weak, surface cyclone. A region of negative EPV and associated CSI was located to the north of the surface low near where the dry conveyor belt overlay the warm, moist air of the trowal airstream. Mid-tropospheric frontogenesis occurred to the northwest of the negative EPV region as the trowal airstream became confluent with the cold air to the north of the cyclone. The snowband formed to the north of the negative EPV zone but to the south of the frontogenesis zone. Moore et al. noted that a gradual southeast-to-northwest transition existed in the stability of the atmosphere from convective instability near the surface, to elevated convective instability, to CSI, to weak symmetric stability. Unlike Nicosia and Grumm (1999) who found that the CCB played an important role in generating a deep, moist layer, Moore et al. (2005) found that the WCB provided the deep moisture along the trowal airstream. Nevertheless the interaction between the conveyor belts, whether for cyclones in the northeastern (Nicosia and Grumm, 1999) or central (Moore et al., 2005) United States, provides a mesoscale region of moisture, lift and instability that is conducive to the formation of snowbands. Others have found similar processes to be important in the development of heavy snowfall-producing bands (e.g. Novak and Horwood, 2002; Novak et al., 2006).

Grim et al. (2007) used high resolution observations to compare the trowal and warm frontal structures of two winter cyclones that produced heavy snow swaths across Illinois, Wisconsin and Michigan. The cyclones had different origins, with one originating over the Colorado Rockies and the other over the Gulf of Mexico. The trowal structure in both of these storm systems was different from the classical trowal structure in which the trowal axis is located at the intersection of the warm front and cold front, the cold front having overtaken and ascended the warm front. In both cases, the movement of the DCB over the warm front resulted in wedging the trowal air mass between the dry air and the warm front. The majority of the heavy snowfall produced by these storms was associated with the precipitation band coincident with the trowal and which
stretched around the north and northwest portions of the cyclone. While both cyclones included a CCB, the CCB did not appear to play a significant role in generating a deep moist layer or in the precipitation production. Rather, the trowal air mass appeared to be the most influential factor in precipitation production, similar to the findings of Martin (1998a,b) and Schultz (2001). While the trowal was bounded by the warm front to the north and an upper-level front to the south, the upper-level front was an upper-level humidity front in the one case and a cold front aloft (see below) in the other.

Finally, Novak et al. (2008) used both observational and model data to investigate the formation of a mesoscale snowband that occurred over the northeastern United States during December 2002. The formation of the band occurred within a region of increasing midlevel frontogenesis due to the sharpening of a midlevel trough, and was characterized by conditional and inertial instability. The mature stage of the band was associated with a significant increase in frontogenetic forcing and an increase in conditional stability in association with the release of the conditional instability. During dissipation the conditional stability continued to strengthen while the frontogenetic forcing decreased. The changes in moisture appeared to play a smaller role than changes in rising motion in the demise of the band. This case differs from a number of the previous snowband studies in that the maximum rising motion was found close to the frontogenesis maximum, as opposed to being located 50 to 200 km away on the warm side of the front. CI was also evident at least 1.5 hours before the band, and the release of this CI through frontogenetical forcing resulted in the formation of the band and increased conditional stability. Traditionally band formation has been associated with decreasing conditional and symmetric stability. Even more recently, simulations conducted by Novak et al. (2009) showed that latent heat released by the band itself was highly important to both the formation and the maintenance of the band.

10.3.5.4. Mesoscale Features of the Shapiro-Keyser Model

The original rainband classification studies were conducted for extratropical cyclones more typical of the Norwegian cyclone model. The distribution of precipitation associated with the features of the Shapiro-Keyser cyclone model will be discussed in this section. Some of the earlier observations of these features came from the examination of extratropical cyclone development during the Experiment on Rapidly Intensifying Cyclones over the Atlantic (ERIC; Hadlock and Kreitzberg, 1988; Wakimoto et al., 1992; Neiman and Shapiro, 1993; Neiman et al., 1993). Much of the analysis focused on the record intensity cyclone that occurred on 4-5 January 1989 over the Atlantic Ocean east of North Carolina shown in Fig. 10.16. As described in detail by Neiman and Shapiro (1993) and summarized here, a baroclinic leaf cloud initially observed at 0000 UTC evolved into a comma-shaped cloud system by 0600 UTC (Fig. 10.16a). Cold cloud tops associated with deep (8-10 km)
FIGURE 10.16  NOAA GOES-East infrared satellite images at (a) 0600 UTC, (b) 1200 UTC, (c) 1800 UTC 4 January 1989, and (d) 0000 UTC 5 January 1989. Surface frontal positions are shown. (After Neiman and Shapiro (1993))

cumulus convection were located parallel and to the east of the cold front, and extended northward into the region of the warm front. The comma head cirrus clouds had grown in extent, and descent of dry stratospheric air to the northwest of the cold front is evident. Also evident at this time was the development of the bent-back front as it extended westward into the polar airstream.

By 1200 UTC (Fig. 10.16b) the T-bone phase became evident and surface cyclogenesis was focused on the bent-back front. The cloud band associated with the cold front to the east formed the tail of the comma head and narrowed in cross-frontal scale. Stratocumulus cloud streaks were evident to the west of the cyclone, and orientated parallel to the north-northwesterly flow of cold, continental air over the cool (∼5 °C) ocean surface. These clouds were initially limited to the shallow boundary layer (∼1 km deep) but developed into deeper cellular cumulus bands (∼3 km) further downstream as this airmass moved over the warmer waters (∼20 °C) of the Gulf Stream. By 1800 UTC (Fig. 10.18c) the bent-back front and the comma-head began to wrap around the cyclone center in association with the development of the warm core seclusion phase, and a mesoscale cloud-free “eye” was evident. The cold-topped clouds of the comma
head tail were situated \( \sim 100 \) km ahead of the surface cold front. The only clouds located at the leading edge of the cold front were located within \( \sim 250 \) km of the triple point, and were relatively deep (>8 km) convective clouds. While the cold front was collocated with a line of intense convection, the causes of the ascent under the cold cloud tops could not be ascertained. Cloud streaks associated with the offshore flow surrounded the western, southern and eastern regions of the cyclone, and vertically-enhanced cloud streaks were located along the secondary cold front where air-sea temperature and moisture differences were greatest. By 0000 UTC (Fig. 10.16d) the low clouds associated with the bent-back front, surrounding the “eye” were still obvious. The eastward displacement of the comma head tail relative to the cold front was also still evident, although the deep convection located along the cold front to the south of the triple point was no longer obvious.

Wakimoto et al. (1992) and Neiman et al. (1993) both made use of airborne radar data to examine the mesoscale cloud and precipitation characteristics of the case just described. Intense mesoconvective precipitation bands extending to between 8 and 10 km were observed along each front, as well as the region of frontal fracture. While deep convection had previously been observed along the cold front (Hobbs and Biswas, 1979; James and Browning, 1979; Hobbs and Persson, 1982; Parsons and Hobbs, 1983), such convective organization had not been observed along the warm front. In addition to the deep convective elements, shallower (<4 km) widespread precipitation was also produced along the warm front, being found in the regions of frontal upglide north of the front, including its bent-back extension. The heaviest precipitation along the cold front was produced by narrow rain bands spaced \( \sim 25-40 \) km apart and orientated at \( \sim 30^\circ \) to the front.

Along the warm front, the southerly ascending airstream of the warm conveyor belt produced shallow (<4 km) upglide precipitation and vertical motions of the order of 2-4 m s\(^{-1}\). The front in these regions was approximately 20 km wide, and had a slope of \( \sim 1:9 \). However, in the mesoconvective elements the warm conveyor belt ascended to \( \sim 10 \) km over a horizontal distance of \( \sim 10 \) km. The reflectivities were of the order of 45-50 dBZ, the cross-frontal convergence collapsed to storm-scale (\( \sim 3 \) km) and the frontal slope was \( \sim 1:1 \), making it much steeper than previous observations of warm fronts (e.g. Locatelli and Hobbs, 1987). The implied ascent along the axis of the conveyor belt in these elements (> 25 m s\(^{-1}\)) is comparable with the vertical motions within mid-latitude continental storms. The ascent of the warm, southerly air over the warm front therefore alternated between the “escalator” of the slantwise frontal upglide in regions of shallow precipitation and the “elevator” within the mesoconvective updrafts, where the convective cloud scale processes influenced the warm front circulations. A schematic showing this escalator-elevator concept is shown in Fig. 10.17.

The convection associated with the bent-back front was of moderate intensity with cloud tops extending to \( \sim 8 \) km. In contrast to the intense, nearly
FIGURE 10.17  Schematic of the “escalator-elevator” perspective of warm-frontal ascent, as the warm conveyor belt (flat, lightly stippled arrows) rises over the cold conveyor belt (tubular dashed arrow). Mesoconvective ascent (the elevator, solid arrows) and convective clouds (stippled with white anvils) are shown at regular intervals between regions of upglide ascent (the escalator). (After Neiman et al. (1993))

continuous lines of convection associated with the cold and warm fronts, the convection was more scattered in nature, occurring along and to the north of the bent-back warm front. During the initial stages of the warm core seclusion phase, the convective elements encircled the western and southern regions of the developing warm core seclusion. In the mature warm seclusion phase, the bent-back front completely encircled the cyclone center, thus secluding a mesoscale region (125 km) of air with high $\theta_e$ (20-30 K higher than on the cold side of the bent-back front). The equivalent potential temperature of this air increased by 8-10 K between 1800 and 0000 UTC, even though it was secluded from the warm sector. This occurred due to an increase in boundary layer moisture resulting from upward latent heat fluxes acting on air parcels spiraling into the warm core seclusion.

In an examination of an extratropical cyclone that developed during IOP5 (19-20 January) of ERICA, Blier and Wakimoto (1995) noted that the strongest radar echoes occurred along the primary cold front and the bent-back to the northeast of the low, and that the widths of the precipitation bands associated with the cold, warm and bent-back fronts were comparable. Also, while precipitation increased in intensity along the bent-back front in the direction toward the low center, there was no convective activity or convection along the secondary cold front that extended beyond the low. Both the bent-back front and the warm front had narrow cross-frontal scales.

Until Martin’s (1998a,b) observational and modeling study of extratropical cyclone development over the central United States during January 1995, the bent-back front had not been identified in continental cyclones. The cyclone was characterized by a deep warm front, a cold front and a bent-back front that showed a similar structure to the bent-back front of the maritime cyclones previously described, although the evolution of this feature may have been different to that observed for the maritime cyclones. In this case, the bent-back front was stronger above the friction layer, suggesting that there may be stronger similarities between continental and maritime cyclones than was previously
thought. Martin also found that this cyclone produced the classical occluded structure, first at upper levels and then at the surface, and concluded that a warm-occluded structure may exist at mid-levels while a fractured frontal structure occurs at the surface, and that at least some cyclones that exhibit warm-occluded structures develop those structures as a result of the seclusion mechanism.

Finally, it should be noted, that these observational studies have demonstrated that the scale of significant frontal and circulation features within the Shapiro-Keyser model, such as the bent-back front and the fracture of the cold front from the warm front, are mesoscale processes. Such considerations should be borne in mind when numerically simulating these systems if the cloud and precipitation processes associated with these features are to be accurately represented.

10.3.5.5. Orographic Influences on Precipitation Bands

When an extratropical cyclone moves over a mountainous region, the horizontal distribution of precipitation typically associated with fronts and rainbands is influenced by orographic forcing, thus making rainband classification in the various cyclone sectors difficult. The precipitation processes of extratropical cyclones tracking over the mountains of the Pacific Northwest have been researched for several decades (e.g. Hobbs et al., 1975; Houze et al., 1976; Marwitz, 1987). Several recent studies have focused specifically on the role of orography in rainband initiation and formation. Yu and Smull (2000) examined a cold frontal system as it made landfall along the mountainous coast of Oregon, and found that upstream blocking by the coastal terrain led to the rapid genesis of a NCFR, and the enhancement of two prefrontal rainbands. The blocking of the low-level prefrontal flow occurred as a result of the acute orientation of the front to the coast. This resulted in significantly intensifying the prefrontal, low-level along-barrier flow and changing the cross-frontal vertical wind shear. Finally, as the front moved southward the along-barrier flow decreased and the NCFR dissipated. Colle et al. (2002) found from their modeling studies of the same case study that rapid development of the NCFR occurred, even in the absence of topography. However, the coastal topography helped to strengthen the thermal gradients through enhanced deformation frontogenesis driven by the terrain-enhanced prefrontal flow, thereby strengthening the vertical circulation and the NCFR.

Braun et al. (1997) investigated the changes in the mesoscale characteristics of an intense frontal system that moved toward the Oregon shoreline on 8 December 1993. As the frontal system approached the shore the low-level prefrontal flow increased while the wind perpendicular to the front decreased, thereby affecting the evolution of the frontal rainband. The offshore NCFR was characterized by deep convection which, together with the lightning frequency, decreased as the frontal system approached the coast, thus suggesting a stabilization of the lower troposphere in this region. Even though the radar echo depths and maximum updrafts decreased as the NCFR moved toward
the coast, the maximum reflectivity at lower levels didn’t change significantly, implying a change in the microphysical characteristics of the frontal bands as the system approached the coast, an observation supported by changes in lightning frequency. The stronger mid-to-upper levels updrafts in the offshore bands appear to have increased the production and upward transport of ice particles to form the cloud and stratiform precipitation observed ahead of the NCFR.

Kingsmill et al. (2006) analyzed ten extratropical cyclones that underwent landfall during the California Landfalling Jets (CALJET) experiment using data from two sites, one in the mountains and one in the Central Valley in northern California. In their 10-case composite they noted a distinct change in the slope of the radar reflectivity about 2.5 km above the brightband that represents a change in the hydrometeor growth rate. This structure, which they call a “shoulder”, remained at approximately 2.5 km above the brightband in the cold sector, warm front, warm sector, cold front and cool sector regimes and was evident at both sites indicating that this feature occurs independently of orographic effects. About one third of the examined profiles did not show the brightband. These non-brightband profiles exhibited an increase in reflectivity with decreasing altitude in the lower levels, suggesting that growth was occurring by collision-coalescence processes. The relationship between the surface rainfall rate and the low-level radar reflectivity implied that a larger number of small drops occurred in the mountainous profiles than would be obtained using a Marshal-Palmer drop size distribution, a trend especially obvious for the non-brightband profiles. The drop size distributions of the valley profiles, on the other hand, more closely followed the Marshall-Palmer size distribution, therefore indicating larger droplets in these profiles than at the mountainous site, and that non-brightband rainfall occurs less frequently in the valley. Finally, the rainfall rates were greatest in the cold-frontal regime, and the non-brightband rainfall was most frequent during the warm-frontal, warm-sector and cool-sector regions.

A synergistic interaction between frontal and orographic processes was suggested by Woods et al. (2005) in their analysis of the passage of a strong extratropical system over the Cascades. In this context, the term synergistic means that the combination of the two processes produces a greater amount of precipitation than if either process were acting alone. The storm was characterized by a lower-tropospheric frontal zone that tipped forward with height, and was overlaid by a backward tilted upper cold front. A broad rainband was located ahead of the surface cold front. The structure was similar to a warm occlusion, however a distinct warm front could not be found and hence their use of the term tipped-forward cold front. Tipped-forward cold fronts have also been identified in other studies (e.g. Locatelli et al., 2002a, 2005). Woods et al. suggested four possible combinations of cross-barrier flow (easterly and westerly) and lower-tropospheric frontal (tipped-forward or tipped-backward) structures (Fig. 10.18). In each case, a mid-to-upper level frontal cloud produces
ice crystals which fall to the lower troposphere. In Fig. 10.18a, the low-level cross-barrier flow is easterly and the frontal structure is backward-tilted. Under this scenario, precipitation rates are not enhanced due to the lack of orographically-generated cloud liquid associated with the easterly downslope flow. The westerly flow behind the front may produce some lower level orographic cloud which in turn may enhance the growth rates of the ice particles falling from the rear of the upper-level frontal cloudband. Following the frontal passage, the precipitation is purely orographic in association with the postfrontal westerlies. The setup in Fig. 10.18b is similar to that in Fig. 10.18a, except that the lower-tropospheric front is now tipped forward. In this case, there is no low-level upslope flow underneath the upper-level precipitation as the prefrontal, downslope easterlies extend all the way to the surface front. Thus any form of enhancement through orographic related processes cannot occur. This situation has not been described in the literature, although tipped-forward fronts with prefrontal easterlies are common in the Pacific Northwest and were observed in the Improvement of Microphysical Parameterization through Observational Verification Experiment (IMPROVE-1) (e.g. Locatelli et al., 2005; Evans et al., 2005).
In Fig. 10.18c, the setup differs from Fig. 10.18a in that the prefrontal flow is now a warm, moist southwesterly flow which produces significant amounts of cloud liquid-water through orographic forcing. However, only some of the upper-level frontal precipitation falls into this cloud. Most of the upper-level precipitation falls behind the surface front into the more shallow, postfrontal orographic clouds. Such a scenario was observed in the 8-9 December 2001 case of IMPROVE-2 (Bond et al., 2005). Finally, the southwesterly, moist cross-barrier flow and the tipped-forward frontal structure observed in Wood et al.’s case study are shown in Fig. 10.18d. This scenario represents the situation that has the greatest potential for producing the most precipitation. The low-level flow produces significant amounts of cloud liquid-water through orographic forcing, and the tipped-forward front results in the entire upper-level frontal precipitation band being located over the lower orographic cloud, thus providing the optimal situation for ice particles from above to seed the lower orographic cloud, thereby removing cloud water that otherwise may not have been removed. Woods et al. (2005) estimated that such liquid-water scavenging clouds result in precipitation enhancements of 0.4 to 1.0 mm hr$^{-1}$. Finally, if the upward motion associated with the orographic forcing is sufficiently deep, it can enhance snow production in the upper-level cloud band (Colle, 2004).

Medina et al. (2007) analyzed 16 extratropical cyclones that crossed the Oregon Cascade Mountains during IMPROVE-2 (Stoelinga, 2003) in November 2001. They examined the mountain-induced variations in precipitation of a storm by focusing on the vertical structures observed at a particular site. They defined three basic sectors of extratropical maritime cyclones: (1) the early sector which is the first to pass over a surface site, is identified with the warm advection region of the cyclone and which often contains a warm front that slopes gradually down toward storm center; (2) the middle sector in which the warm advection transitions to cold advection, the warm front often reaches the surface, and cold and occluded frontal structures may occur; and (3) the late sector which is the cold, unstable zone behind the cold and/or occluded front and is often called the postfrontal period. The radar echo structures associated with each of these stages are shown in Fig. 10.19.

The early sector of the storm’s passage over the windward slopes is characterized by a leading edge echo (LEE) in which a deep layer of precipitation appears aloft initially, and then descends toward the surface until a deep stratiform echo extends from the surface to $\sim$6 to 7 km (Fig. 10.19a). Updraft cells may be embedded in the LEE at upper levels. The LEE period echoes do not appear to be qualitatively influenced by the orography. The precipitation over the windward slope within the middle sector of the storm extends in a vertically continuous column from the mountainside up to a height of about 5-6 km (Fig. 10.19b). The vertical structure of the echo is one of a double maximum echo (DME) in which the lower echo is the melting-generated brightband, and the second region of high reflectivity is located 1 to 2.5 km above the brightband. The latter is not evident when the middle sector of the
storm passes over the regions upwind of the Cascades but becomes obvious over the windward slope, apparently as a result of the dynamic interaction of the storm with the terrain. A layer of turbulent overturning exists between the two reflectivity maxima with small-scale updraft cells \( (w > 0.5 \text{ m s}^{-1}) \) that may play an important role in the precipitation enhancement on the windward slope. The late sector is characterized by generally isolated shallow convection echoes (SCEs) (Fig. 10.19c). These cells that are observed in the late sector
of extratropical cyclones offshore of Oregon and Washington become broader on the windward slope, probably in response to orographic uplift. The shallow convective echoes have top heights lower than those observed during the LEE and DME periods, and disappear more rapidly on the lee side of the Cascades compared to the lee side precipitation of the LEE and DME periods.

10.3.5.6. Precipitation Cores and Gaps

Radar reflectivity has shown that the rainfall within NCFRs is typically not uniform but rather is organized into alternating cores of precipitation and gap regions (Hobbs and Biswas, 1979; James and Browning, 1979; Locatelli et al., 2005; Wakimoto and Bosart, 2000). The precipitation cores tend to be regularly spaced ellipsoidal cells orientated at an angle to the surface cold front, although their shapes and orientation to the front are not always uniform, even within the same front. While precipitation cores have also been observed within WCFRs, they tend to be irregularly shaped without any distinct organization to them (Moore, 1985). Interactions between the cores depend on the relative strength of each core and how close they are to one another. The gap regions in NCFRs tend to vary in size, with gaps greater than 10-12 km occurring less than 20% of the time and being referred to as “large” gaps by Locatelli et al. (1995). The large gaps are associated with weak surface wind discontinuities, lack strong updrafts and appear to be dynamically different from smaller gaps. Gaps have been found to influence the evolution of the cores, and may move along the front at speeds slower than that of the cores.

Several mechanisms have been suggested regarding the formation and control of the precipitation core and gap structure of the NCFR, including horizontal shear instability along the cold front (Hobbs and Persson, 1982; Moore, 1985; Wakimoto and Bosart, 2000; Jorgensen et al., 2003), trapped gravity waves (Brown et al., 1999), and differential advection of precipitation (Locatelli et al., 1995; Brown et al., 1999). To further investigate the horizontal shear hypothesis Moore (1985) performed a linear stability analysis on an idealized frontal zone that consisted of a line of convection coincident with a line of cyclonic shear. The air within the rainband was assumed to be unstable. Moore found that the presence of the horizontal wind shear resulted in a mode with a short wave cutoff, that the coupling between the convective processes and shear instability in this mode was strong, and that its most unstable wave has characteristics similar to the precipitation cores in NCFRs. Moore also noted that as the shear increased, the cells become elongated and narrow.

Locatelli et al. (1995) observed that spacing between the precipitation cores, the height of the head of the density current associated with the cold front, and the strength of the cold front were all positively correlated to the precipitation strength, thus suggesting that the precipitation cores are strongly influenced by precipitation. Based on their observations they developed a new conceptual model. In their model, while the formation of precipitation cores may initially be controlled by processes other than precipitation processes, the latter soon
become dominant. As the precipitation cores develop, diabatic frontogenetic processes associated with precipitation evaporation result in an increase in the height of the hydraulic head in the regions of the cores. The hydraulic head moves at a speed that is predicted well by density current theory. The greater the height of the head, and hence the stronger the frontal zone, the greater the speed of the hydraulic head and the associated cores. This results in a positive feedback in that, the greater the speed of the hydraulic head, the stronger the updraft and the associated precipitation. Enhanced precipitation results in increased hydraulic head heights and local enhancements of frontal strength. This in turn is associated with local increases in vorticity (through greater convergence and vertical velocity) within the cores, and hence a perturbation in the frontal boundary. This model has several implications including, that stronger cores move faster than weaker cores, that stronger cores can overwhelm weaker cores, and that fronts will be strongest where the cores are located.

Browning and Roberts (1996) have observed that classical cold fronts and split cold fronts (see section below) can change from one to the other either in space or in time, and that such a change may influence the precipitation structure along the frontal boundary. A common scenario is one in which the cold front is a split front just south of the center of a developing extratropical cyclone, but transitions to a classical cold front further away from the cyclone, such as occurred over the United Kingdom on 8 December 1994. On this day, a dry intrusion moved over a tongue of relatively moist air and advanced ahead of the surface cold front near the center of the cyclone, thus forming a split front structure, while to the south of this, along the trailing portion of the cold front, the dry air undercut the moist air ahead of the front in a classical cold front structure. While line convection occurred along the entire NCFR, the almost continuous line convection in the region of the classical cold front structure transitioned to distinct, broken convective elements that were orientated at approximately 30° or more to the surface cold front in the split front region, as shown in Fig. 10.20. Individual convective elements developed at the northern end of the classical part of the front and advected to the split front part of the cold front, where they were tracked for over 4 hours (Fig. 10.20a). The transition to the broken line appears to have been primarily associated with a change in the orientation of the surface cold front with respect to the preferred orientation of the convective line elements, and only partly due to a change in the orientation of the line elements themselves (Fig. 10.20b). This mechanism therefore appears to be different from shearing mechanisms suggested previously in which the distinct convective elements result from the reorientation of sections of the convective line as a result of shearing instabilities.

The output of a 3-D numerical simulation of an observed NCFR conducted by Brown et al. (1999) demonstrated precipitation cores approximately 40 km in length and rotated counterclockwise from the cold front, results that were
consistent with observations. Brown et al. found that convection associated with each precipitation core generated a gravity wave, similar to the scenario in which air flows over an isolated mountain peak. These gravity waves are V-shaped, with the apex of the V orientated slightly counterclockwise to the cold front. The eastward branch of the V, or that branch closest to the cold front was
found to extend over the cold front. The motion of the first wave oscillation is downward in direction, strongest in magnitude, and results in warming aloft thereby producing a gap region. The next oscillation of the gravity wave is weaker, upward in direction and induces convection further north. In this manner, trapped gravity waves produce a series of precipitation cores and gaps. Brown et al. also noticed that the elliptical shapes and orientations of the precipitation cores appear to be due to the advection of precipitation by the core-relative winds.

Wakimoto and Bosart (2000) conducted a detailed analysis of airborne Doppler radar data of an oceanic cold front exhibiting precipitation core and gap regions. They found from their observations that the precipitation cores were due to both horizontal shearing instability and the advection of hydrometeors by the precipitation core-relative winds. Also, unlike previous work, a strong discontinuity did not exist along the whole length of the precipitation core. While the southern regions of the precipitation cores may be associated with a discontinuity and lighter precipitation, the variables changed more slowly in the northern regions of the cores and the precipitation was more intense. The cold front in the regions of the precipitation cores appeared to propagate as a density front, although density current theory did not appear applicable to the overall motion of the cold front.

Many of the features previously observed to occur with NCFRs were evident in a study by Jorgensen et al. (2003) of an intense NCFR observed using airborne Doppler radar during the Pacific Coastal Jet Experiment, and simulated using MM5. These include gaps in the narrow precipitation zone, a strong frontal updraft at the leading edge of the cold air, strong shear in the low levels of the inflowing air, a low-level jet parallel to the NCFR with a maximum near 1.5 km AGL, and a rear-inflow jet of westerly air peaking at \( \sim 1 \) km AGL behind the front. Like Wakimoto and Bosart (2000), Jorgensen et al. (2003) also noticed differences between the northern and southern regions of the precipitation cores within the NCFR. In particular they found that differences in the updraft tilt exist consistently on either side of the NCFR gap regions (Fig. 10.21). Within the precipitation core the updrafts tend to be erect (Fig. 10.21 A-A') with narrow rainfall regions, representing the optimal vertical shear balance (Rotunno et al., 1988). Near the south end of the precipitation core (Fig. 10.21 B-B') the environmental shear is weak relative to the cold air circulation, thus causing the updraft to tilt upshear. This region is associated with a broader zone of precipitation. Finally, at the northern end of the precipitation core, the environmental low-level shear is stronger and the updraft tilts downshear with height (Fig. 10.21 C-C'). Rain is this region falls ahead of the cold air. It is thus the interaction of the density current and the ambient cross-frontal vertical wind shear, rather than the density current dynamics alone, that explains the updraft and precipitation characteristics. The prefrontal cross line flow shows the greatest variation along the NCFR, while the depth and strength of the cold air varies little across the gaps. The gaps will not be self-sustained, as
convection to the south of the gap will weaken with time compared with its northern counterpart, due to evaporative cooling as the inflow air passes through the rainshaft. The observations showed that the gaps only lasted for several hours.

10.3.5.7. Split Fronts and Cold Fronts Aloft

The limitations of the Norwegian cyclone model in explaining all of the observed precipitation distributions of extratropical cyclones, particularly within the warm sector (such as squall lines that form at least 200 km ahead of surface cold fronts), led to the identification of split fronts (e.g. Harrold, 1973; Browning and Monk, 1982). More recently, the recognition of the influence of topography on frontal structures in the central United States resulted in investigations into cold fronts aloft (CFA) (e.g. Hobbs et al., 1990; Locatelli et al., 1995, 1998, 2002b; Stoelinga et al., 2000), and the
subsequent development of a more general conceptual model for cyclones in
this region referred to as the Structurally Transformed by Orography Model
(STORM) (Hobbs et al., 1996). As Locatelli et al. (2002b) pointed out, cold
fronts aloft (where by this we are generically referring to baroclinic zones
above the surface that are significantly ahead of those at the surface) and
their associated wide cold frontal rainbands form an important part of the
classical model’s warm occlusion (Bjerknes and Solberg, 1922), the trowal
model (Galloway, 1958), the split front model (Browning and Monk, 1982)
and the STORM model (Hobbs et al., 1996). The common feature of the cold front
aloft in all four of these models, even though called by different names (upper
cold front in the split front model and cold front aloft in the warm occlusion
and STORM model) is that it is a baroclinic zone associated with a thermally-
direct circulation. These cold fronts aloft differ from those upper level fronts
previously described (e.g. Keyser and Shapiro, 1986), which are more closely
associated with dynamical processes near the tropopause. The role and forcing
specifically associated with the orography are addressed in Chapter 11, whereas
the mesoscale and precipitation characteristics of cold fronts aloft are addressed
here.

Browning and Monk (1982) studied nine cold fronts associated with
cyclones crossing the United Kingdom in order to develop a simple model of
split fronts. In this split front model the “upper cold front” is located up to
several hundred kilometers ahead of the surface cold front as a result, either of
overrunning the surface cold front, or of the modification of the low-level air
(Fig. 10.22a). The warm conveyor belt extends ahead of the surface and upper
cold fronts and above the surface-based warm front. Low-θ_e air within the dry
conveyor belt overruns the warm conveyor belt (Fig. 10.22a), thereby creating
potential instability. When this potential instability is released in association
with the upper cold front, a convective rainband is produced ahead of the surface
cold front. As the classical cyclone model does not distinguish between upper
and surface cold fronts, it cannot describe the rainbands associated with these
split fronts. The band itself is often observed to extend from within the warm
sector, where the precipitation associated with the upper cold front tends to
be more intense and cellular in character, to beyond the warm front where the
precipitation is produced by upgliding over the warm front and hence is more
frontal in nature.

Behind the upper cold front a sharp drop in cloud cover is typically observed
(although bands may be evident), followed by a shallow moist zone made up of
warm conveyor belt air (2-3 km deep) located between the upper cold front
and the surface cold front (Fig. 10.22b). Uplift of the warm conveyor belt
air by weak convection may produce patches of light rain and drizzle in this
region, although precipitation is frequently suppressed due to the deep layer of
dry air aloft that limits the depth of the moist layer. Widespread stratus may
also develop after the rainfall produced by the upper cold front has passed, the
base of which tends to lower until lifted by the passage of the surface cold
front. The surface cold front tends to produce further light rain and drizzle, although occasionally a narrow band of convection develops. While changes in temperature across the upper and surface fronts tend to be small due to the subsidence of air behind the front, the drop in humidity is significant. Finally, split fronts can be distinguished from trowels by the presence of unoccluded warm air between the upper and lower fronts.

FIGURE 10.22 (a) Schematic plan view of a split front (from Koch, 2001, after Browning and Monk, 1982). The broad arrow depicts the warm conveyor belt (high-$\theta_w$ air) which gently rises in an isentropic coordinate systems relative to the moving cyclone. The narrow arrows represent the flow of low-$\theta_w$ air within the dry conveyor belt, which descends while overrunning the warm conveyor belt. The leading edge of the low-$\theta_w$ air marks the position of the split cold front (open symbols) just ahead of which appears a rainband. Surface fronts are shown by standard symbols. (b) Vertical cross section along A-B showing (1) regions of light precipitation along the warm front, (2) intense precipitation produced by the split-front rainband, and (3) shallow, light precipitation occurring in the warm sector ahead of the surface cold front. The dry conveyor belt is depicted by the broad arrow behind the split front. (After Koch (2001))
Hobbs et al. (1990) proposed the cold front aloft model in order to explain frontal features often observed to the east of the Rocky Mountains. In this model there is a warm or Arctic front at the surface, a CFA and its associated short wave aloft, and a surface trough located 200-300 km behind the leading edge of the CFA. The surface trough may take the form of a cold front, a warm occluded front, a lee trough, a lee trough/Arctic trough combination, or a dry trough (a lee trough that has the characteristics of a dryline) as described in the STORM conceptual model (Hobbs et al., 1996). When the surface trough is a cold front, the CFA model is equivalent to the split cold front model of Browning and Monk (1982). In the CFA model, the air to the east of surface trough is characterized by high-$\theta_e$ values. As dry descending air from over the mountains overlays this warm moist air to the east of the trough, potentially unstable conditions develop. The upward motion then produced by the passage of the CFA can lift this potentially unstable air thereby producing a CFA rainband. This rainband can be composed of several squall lines and may be associated with tornadoes, hail, and flash flooding over several states. Hobbs et al. (1990) proposed adding squall lines that form in association with CFA to the two previously identified squall line types (prefrontal and ordinary). As the precipitation produced by these events is associated with the leading edge of the CFA, the surface precipitation falls well to the east of the surface trough, as it does in the split front model.

The main difference between the split front and the CFA is that the split front is an upper cold front located over the warm sector, ahead of the surface cold front (Fig. 10.22), whereas the CFA is ahead of a surface trough. Thus, as stated by Locatelli et al. (1995), in the CFA model “the region analogous to the classic warm sector of a warm occlusion is located west of the surface position of the lee trough, not east of it as required by the split-front model.” Also, the CFA forms as a result of orographic blocking of the lower portion of an advancing cold front and subsequent adiabatic warming by the descent of the flow over the lee slope of the mountains, both of which act to destroy the cold front near the surface. In the split front model, which was developed based on case studies observed over the United Kingdom, no such topographical influences were necessary.

Following their original CFA study (Hobbs et al., 1990), and based on a number of subsequent studies by Locatelli et al. (1989, 1995), Martin et al. (1990, 1995) and Wang et al. (1995), and Hobbs et al. (1996) went on to develop a more generalized conceptual model of cold fronts aloft. The model incorporates a number of features including a dry trough, an Arctic front, a low-level jet, and two rainbands, the CFA rainband (Locatelli et al., 1995, 1998) and the pre-dry-trough rainband (Martin et al., 1995). Both of these rainbands can produce heavy precipitation and severe weather ahead of the dry trough. The aspects of this conceptual model are described in detail in Hobbs et al. (1996) and only briefly summarized here. As a shortwave trough moves eastward, westerly downslope flow over the Rockies increases, producing adiabatic warming and a lee trough. Confluence of warm, dry air off the Rockies and warm, moist air from the Gulf of Mexico produces a west-east moisture
gradient, and a trough with the characteristics of a dryline may develop called the dry trough. This dry trough has warm front-like circulations that can lift potentially unstable air. The moist air flowing northward from the Gulf rises and turns toward the northeast as it approaches the dry trough. The dry, downslope air off the Rockies reaches its lowest point over the dry trough and then rises above the warm, moist Gulf air. In late winter, well-mixed heated air off the Mexican plateau may flow eastward above these two airstreams.

As the dry trough develops, the strong southerly flow ahead of the dry trough enhances the formation of a southerly LLJ jet which in turn enhances the transport of warm, moist air northward. The airflow associated with the dry trough results in dry warm (low-$\theta_e$) air being superimposed over warm moist (high-$\theta_e$) air, thereby producing a potentially unstable environment ($\theta_e$ is decreasing with height), which when lifted becomes unstable and produces convective clouds and precipitation (Fig. 10.23a). The resultant convective rainband is referred to as the pre-dry-trough rainband (Martin et al., 1995) (Fig. 10.23b). The pre-dry-trough rainband typically moves northward and eastward away from the dry trough, may be composed of several sub-bands, is often associated with severe weather and may extend over many states. The southward movement of the Arctic front and the lifting of warm moist air around the associated low pressure center, produces precipitation that is often in the form of freezing rain or snow (Stoelinga et al., 2000), and which can at times be heavy should the flow be suitably located to be upslope flow. Thus precipitation that is both stratiform (Arctic front) and convective (pre-dry-trough) may occur simultaneously.

The most important feature of the STORM model is the rainband that forms in association with the CFA, called the CFA rainband (Fig. 10.23c). A surface front may accompany a CFA as it moves over the Rockies, although the surface front tends to become eroded with the adiabatic warming associated with the downslope motion of the front. As the CFA moves over the dry trough it occludes with the high-$\theta_e$ air ahead of the dry trough in a structure typical of warm occlusions. Upward motion produced by the CFA can then lift the potentially unstable air to the east (ahead) of the dry trough, producing what is called the CFA rainband. This rainband may consist of several squall lines, and can produce severe weather including tornadoes, large hailstones and flash floods, from the Rockies to the east coast of the United States (e.g. Locatelli et al., 1989, 1995; Sienkiewicz et al., 1989; Hobbs et al., 1990; Martin et al., 1995; Stoelinga et al., 2000). Finally, Hobbs et al. do point out that while they are not proposing that all of the cyclones that develop to the lee of the Rockies fit their model, many of the cyclones that do develop in winter, spring and even early summer appear to be better described by this model than the classic cyclone model. The use of the classic model would result in these rainbands and their associated severe weather not being forecast.

The impacts of cold fronts aloft (in the generic sense) on various mesoscale systems and processes continue to receive attention. The precipitation produced
in association with cold fronts in the upper air can be nonconvective, weakly convective or severely convective, being dependent on the stability of the atmosphere being lifted by the cold front. The structure of cold fronts aloft is ideal for supporting squall lines in that the cool, dry air behind the nose of the front supplies the rear inflow air that is required to evaporate precipitation, thereby maintaining the squall line (Locatelli et al., 1995). A positive feedback
mechanism can even exist between the CFA and the squall line, in that the evaporative cooling behind the CFA and the latent heat release ahead of it, associated with the squall line development, strengthens the CFA, which in turn strengthens the squall line and so on. Locatelli et al. (1998) observed a squall line embedded within a rainband associated with a CFA, although the structure was quite different from the typical leading line/trailing stratiform squall line model. They concluded that the CFA, rather than the lower-level cold-pool, was responsible for the squall line generation and maintenance. Geerts and Hobbs (1991) noticed in their observations of a rainband associated with a CFA that the CFA interacted with the boundary layer not only through evaporative processes but also by a shallow downdraft to the rear of the rainband’s rainshaft. Locatelli and Hobbs (1995) reanalyzed a record rainfall-producing storm that occurred over Holt, Missouri on 22 June 1947. This storm produced approximately 1 foot of rain within 42 minutes. It was previously analyzed by Lott (1954) who concluded that the storm was the result of local intensification of a warm sector convective storm a short distance ahead of the surface cold front. Locatelli and Hobbs demonstrated that the STORM conceptual model more accurately described this storm.

More recently, Brennan et al. (2003) made use of a numerical model to examine the impacts of a split front rainband on cold air damming (CAD) to the east of the Appalachian mountains. They found from their simulations that, as the rainband moved over the cold dome, the vertically-integrated latent heat released by the rainband resulted in low-level pressure falls, isallobaric convergence within the cold dome and the resultant inland surge of the coastal front, which is synonymous with the retreat of the CAD cold dome. It should be noted that the cold dome was nearly saturated before the split front passed over the region, thereby reducing any evaporative cooling effects. Thus, should a saturated dome begin to erode before the passage of a split front rainband, the presence of the rainband would not prevent this erosion, and may in fact accelerate the process. This offers an alternative view to the idea that precipitation associated with the rainband would act to strengthen the CAD. Businger and Baik (1991) also examined the impacts of a CFA on CAD and found that the significant source of instability associated with the CFA allowed for the development of severe convection above the cold and highly stable CAD air mass.

More recently, Han et al. (2009) noted a synergistic interaction within an intense cold-frontal rainband between a low- and upper-level front that led to variations in the updraft structure and ice generation along the front. The low-level front was associated with a NCFR that produced heavy precipitation, while the upper-level front was associated with a WCFR that was located either behind or in conjunction with the NCFR along its central and northern regions, and that generated stratiform rain. In the central regions of the front, the link between lifting along the NCFR and lifting associated with the upper-level front resulted in a zone of deep upward motion that produced snow and graupel, while those
regions to the north and south produced relatively little snow and graupel as ascent associated with the upper-level front did not occur in these regions.

10.3.5.8. Gravity Wave Influences

Mesoscale gravity wave disturbances typically have durations greater than 4 hours, periods between 0.5 and 4 hours, wavelengths of 30 to 400 km, and 2 mb pressure perturbations, although pressure jumps of up to 11 mb in 15 minutes (Schneider, 1990) and wind gusts up to 68 kts have been observed. They have been associated with many mesoscale weather phenomena including squall lines, heavy snowstorms, precipitation bands and thunderstorm initiation (Uccellini and Koch, 1987; Bosart and Seimon, 1988; Feretti et al., 1988; Koch and Golus, 1988; Schneider, 1990). These wave disturbances have been observed as both wave packets (Bosart and Sanders, 1986) and singular waves (Ramamurthy et al., 1990), and various mechanisms have been proposed to explain the formation of the gravity waves including geostrophic adjustment (e.g. Koch and Dorian, 1988; Bosart and Seimon, 1988; Ramamurthy et al., 1993), convection (e.g. Bosart and Cussen, 1973; Uccellini, 1975), vertical wind shearing instability (Einaudi and Lalas, 1973; Stobie et al., 1983), and triggering by the leading edge of downslope flow (e.g. Karyampudi et al., 1995; Rauber et al., 2001). It has also been argued that gravity waves initiate convection, rather than convection initiating gravity waves, or that gravity waves and convection are mutually dependent (e.g. Koch and Golus, 1988; Powers and Reed, 1993). The occurrence of mesoscale gravity waves with extratropical cyclones has been observed to enhance the conditions produced by the cyclone, strengthen the cyclone itself, and contribute to the formation of precipitation bands within the storm system. Due to their small scale, clear weather and severe blizzard conditions may be separated by only several kilometers.

An excellent case of gravity wave development in association with a rapidly developing extratropical cyclone occurred over the Midwest and lower Great Lakes on 15 December 1987. This case has been extensively examined using both observational data and numerical output. The waves lasted over 10 hours, produced pressure falls of up to 11 mb in 15 minutes, had horizontal wavelengths between 100 and 200 km, propagation speeds of 30 m s$^{-1}$, and produced cloud-to-ground lightning and localized heavy snow (Schneider, 1990). While the heavy snowfall produced by the cyclone was correctly forecast, blizzard conditions associated with the wave disturbances were not well predicted. The wave disturbances significantly influenced the onset time of the heavy snowfall, and the impacts of the waves on the surface pressure and wind fields made identification of the surface cyclone low difficult, thereby further complicating the forecast. In addition to their remarkable magnitude, these waves appeared to interact with the extratropical cyclone itself (Schneider, 1990). The largest amplitude wave propagated through the cyclone center during its rapid intensification phase, and contributed to a 7 mb pressure fall within one
hour. The wave at this stage had a surface pressure minimum lower than that of the cyclone itself.

Schneider (1990) examined previously suggested mechanisms for mesoscale wave formation in this case, including geostrophic adjustment, convection and vertical wind shear. He found that the atmosphere in which the waves propagated was suited to atmospheric wave ducting, which reduces the vertical propagation of gravity waves, and hence enhances long-lived waves (Lindzen and Tung, 1976). To maintain such long-lived gravity waves requires a deep, stable layer in the lower levels, overlain by a conditionally unstable layer above, that will reflect the waves, hence providing a wave duct. A dynamically unstable critical level characterized by vertical wind shear that is capable of generating and/or maintaining gravity wave energy was also present within the conditionally unstable layer. While the growth and maintenance of such waves seemed to be supported by wave ducting, the mechanisms responsible for forming these waves were less apparent. Schneider suggested that, either a mass-momentum adjustment (Uccellini and Koch, 1987) associated with a shortwave in the region, or convectively induced subsidence zones, may have played a role in the wave generation and/or amplification. Gravity wave generation by vertical shear did not appear to be important, as even though vertical shear was present in the conditionally unstable layer, the region covered by the wave duct was significantly more extensive than the localized gravity wave response.

Powers and Reed (1993) were the first to conduct numerical simulations of the 15 December 1987 case in order to examine the dynamics of this system, in particular the roles played by wave ducting, wave-CISK, convection, shearing instability and geostrophic adjustment in the maintenance of these wave disturbances. While the location and time of occurrence of the simulated waves in their control run were similar to those observed, the speed of the model waves was significantly greater than that of the observed waves. A sensitivity test in which the latent heat of condensation was turned off did not produce strong mesoscale gravity waves. Based on observational and modeling data, Powers and Reed suggested that the model waves were generated by convection on the mesoscale and that they were maintained and enhanced by wave-CISK processes, while the observed waves were maintained by both wave-CISK processes and an environment supportive of ducting processes. Calculations of the duct efficiencies showed that the duct was not perfect, and hence some of the wave energy would propagate vertically out of the duct. A further energy source would therefore be required to maintain the wave for several hours. Wave-CISK processes could supply this energy, a suggestion supported by accounts of thunder as the wave disturbance passed by. Powers and Reed also suggested while convection was primarily responsible for the generation of the observed waves and the lack of latent heating was the limiting factor in wave development, that wind shear (e.g. Stobie et al., 1983) may also have played
a relevant role. Geostrophic adjustment (Uccellini and Koch, 1987; Koch and Dorian, 1988) did not however, appear to be important.

Pokrandt et al. (1996) analyzed several weaknesses with the previous studies of the 15 December 1987 case and based on observational and modeling evidence suggested a mechanism different from convection for the formation of these gravity waves. They pointed out that the meso-β scale of the waves was not addressed, and that convection in the atmosphere usually forms on the meso-γ scale, unless other factors organize it on the meso-β scale. Infrared satellite imagery and radar data from the genesis region just before the waves formed suggest that convection was not the primary feature at this time. Instead, a meso-β scale comma-shaped cloud present in this region appeared to develop into the wave disturbances. The comma-shaped cloud appeared to originally be associated with the left exit region of an approaching subtropical jet. Through the use of both observational and model data, Pokrandt et al. developed a new hypothesis for the explanation of the observed gravity waves. The rising motion of the transverse circulation of the jet streak produced a band of enhanced cloudiness and precipitation that moved at the same speed as the jet streak. The transverse circulation also transported potential vorticity from the stable, cold low-level air to the mid-levels, where it formed thin, PV anomalies on the meso-β scale. This circulation resulted in frontogenesis at midlevels. A wave duct was also present. With the spinup of the large-scale circulation as it moved into the duct region, the midlevel PV anomaly was rotated and stretched, thereby generating small-scale vertical motion perturbations within the duct. The waves then became stronger through wave-CISK processes.

A long-lived gravity wave event that occurred on 14-15 February 1992 during the Storm-scale Operational and Research Meteorology-Fronts Experiment Systems Test (STORM-FEST) has been studied by numerous researchers (e.g. Jin and Koch, 1998; Jewett et al., 1999; Trexler and Koch, 2000). Using observations, Rauber et al. (2001) observed that the gravity wave originated at the leading edge of a dry air mass that was associated with downslope flow east of the Rocky Mountains. The gravity wave and a weak rainband were generated simultaneously just behind the leading edge of this dry air mass as the dry air mass ascended over a warm front to the east of a lee cyclone. In the second paper in this series, Yang et al. (2001) noted that as the dry air mass moved over the denser, cold air below the warm front it caused a wave perturbation in the dense fluid, thus forming the mesoscale gravity wave. The air above the warm-frontal inversion was stable to parcel ascent, but close to neutral for lifting of the layer. As the dry air ascended over the frontal inversion, sufficient lifting occurred to generate a weak convective updraft and rainband. Rain produced by the rainband fell into the dry environment behind the leading edge of the dry air mass. Evaporative cooling resulted in descending air and a depression of the inversion height. Using the numerical experiments with and without evaporation processes, Jewett et al. (2003) demonstrated that evaporatively-generated downdrafts produced by the rainband appeared to be
FIGURE 10.24 A GOES visible satellite image of small comma cloud (see arrow) located at the west of a synoptic-scale frontal cyclone. The length of the arrow is approximately 400 km. The West Coast of the United States is visible on the right-hand side of the image. (After Businger and Reed, 1989a; from Reed, 1979)

important to wave genesis. The evaporative processes resulted in the depression of the inversion across the warm front. This in turn produced surface pressure falls and the initiation of the gravity wave.

10.3.6. Polar Lows

Meoscale features other than cloud bands may also be associated with extratropical cyclones, such as that shown in Fig. 10.24 (Businger and Reed, 1989a). This system is an example of a mesoscale cyclonic system that frequently occurs in polar airstreams behind or poleward of cold fronts in the north Pacific, North Sea, north Atlantic (particularly south of Iceland), and other locations where cold air flows over relatively warm water. These subsynoptic scale cyclones that develop poleward of the polar front and jet stream are called polar lows. Although often only a trough of low pressure occurs at the surface, some polar lows do develop closed cyclonic circulations.

Polar lows appear to be associated either with a comma-shaped or a spiral-shaped cloud structure (Rasmussen, 1983). The spiral cloud pattern of a polar low developing over the Bering Sea is shown in Fig. 10.25. Polar lows associated with the spiral-shaped cloud pattern (e.g. Ernst and Matson, 1983; Rasmussen and Zick, 1987) typically form in the cold air away from the polar front and its associated baroclinic forcing (Rasmussen, 1979, 1983). They appear to develop when cold air in the form of a trough or vortex moves over relatively warm water, thereby generating deep convection through CISK. These
Chapter 10  Extratropical Cyclones and High Clouds

FIGURE 10.25 A NOAA-5 infrared satellite photograph of a polar low and cloud streets over the Bering Sea at 2100 UTC 8 March 1977. SNP indicated the location of the rawinsonde station at St. Paul Island. (After Businger and Reed, 1989a)

lows are referred to as polar lows, Arctic lows, Arctic instability lows, or spiral form lows. They are most frequently observed in regions near the edges of ice sheets or ice covered surfaces such as over the Gulf of Alaska and the Bering Sea (e.g. Businger, 1987), the coast of Norway (e.g. Lystad, 1986), and off the coasts of Antarctica (e.g. Carleton, 1979; Carleton and Carpenter, 1989, 1990).

The polar lows associated with a comma-shaped cloud pattern (Fig. 10.24) tend to be larger than their spiral form counterparts, and develop closer to the polar front, often just poleward of a pre-existing frontal boundary, in regions of positive vorticity advection at mid-tropospheric levels (Reed, 1979; Mullen, 1983; Reed and Blier, 1986a,b; Businger and Walter, 1988). The term comma-shaped cloud pattern is often abbreviated to comma cloud, which should not be confused with the comma cloud that develops in association with extratropical cyclones. Frontal structures appear to accompany these comma-shaped polar lows, although on a smaller scale than their extratropical cyclone counterparts. Baroclinicity seems to play a more important role than CISK in these systems. These comma-shaped polar lows may occur in the North Atlantic (e.g. Carleton, 1985), the North Pacific (e.g. Reed and Blier, 1986a,b; Businger and Hobbs, 1987), and the Southern Oceans (Carleton, 1979).
Polar lows are warm core vortices, which resemble tropical cyclones in their structure including a clear “eye” surrounded by deep convection, and can be accompanied by light up to hurricane force winds (Rasmussen, 1979; Reed, 1979; Locatelli et al., 1982; Rasmussen, 1981, 1983; Mayengon, 1984; Forbes and Lottes, 1985; Shapiro et al., 1987). They range in size from several hundred kilometers to approximately 1000 km in diameter, their size distinguishing them from extratropical cyclones. These systems have been observed to produce severe storms including tornadoes (e.g. Reed and Blier, 1986b), and heavy rain and snowfall (Harrold and Browning, 1969). The most favorable environments within which polar lows tend to develop are characterized by cyclonic flow or shear (Reed, 1979; Mullen, 1979), strong heat and moisture fluxes from the surface, typically the ocean (e.g. Reed and Blier, 1986b; Shapiro et al., 1987), and neutral or unstable conditions in the boundary layer, with conditionally unstable conditions up to the middle or upper troposphere (e.g. Rasmussen, 1977; Mullen, 1979; Businger, 1987). Thus they are most frequently found in areas with large temperature contrasts such as in the region of the polar front or at the interface of relatively warm water and cold ice sheets. In developing a seven year winter polar low climatology of the North Pacific Ocean, Yarnal and Henderson (1989) found that early season polar lows were located further north in association with the land or ice edge, while midseason polar lows were found to occur further south away from the land as the polar vortex expands. The polar low frequency decreases late in the season as the polar vortex breaks down.

Polar lows form on a small synoptic or subsynoptic scale and often intensify rapidly (e.g. Rasmussen, 1985; Shapiro et al., 1987; Rabbe, 1987). The temporal and spatial scale of these systems, and the fact that they form in data-sparse regions, makes them difficult to forecast, although recent advancements in satellite technology and increased forecast model resolutions have helped. Research into polar lows has been conducted since the 1960s and 1970s (e.g. Lyall, 1972; Harrold and Browning, 1969; Mansfield, 1974; Rasmussen, 1977, 1979, 1981). Businger and Reed (1989b) provide an excellent in-depth overview of the polar low research conducted until this time. Even though several field campaigns, such as the Arctic Cyclone Experiment (ACE, Shapiro and Fedor, 1986) and the Lofotes cyclone experiment (LOFZY 2005) have been conducted, in situ data of these systems are rare due to their time and spatial scales, with studies by Shapiro et al. (1987) and Douglas et al. (1995) being the exception rather than the rule.

A number of more recent investigations have made use of satellite data in their investigations of polar lows. Lieder and Heinemann (1999) used AVHRR data and ERS and SSM/I retrievals over the Antarctic region of the Southern Pacific and observed that the polar low of interest developed through baroclinic forcing associated with an upper-level trough. Moore and Vonder Haar (2003) used data from the Advanced Microwave Sounding Unit (AMSU) to examine a polar low that occurred over the Labrador Sea on 17-18 March, 2000 and found that the warm core structure of the polar low was clearly identifiable, being...
2-3 K warmer than the background environment. This could be used to track storm motion. Blechschmidt (2008) utilized thermal infrared satellite imagery and satellite derived wind speeds to develop a two-year polar low climatology of 90 polar low events for the Nordic seas.

Numerous modeling studies of polar lows have been conducted over a variety of regions including Hudson Bay (e.g. Albright and Reed, 1995), the Labrador sea (e.g. Pagowski and Moore, 2001), the Denmark Strait region (e.g. Sardie and Warner, 1985), the north coast of Norway (e.g. Nordeng and Rasmussen, 1992), the Bering Sea (e.g. Bresch et al., 1997) and the Japan Sea (e.g. Yanase et al., 2004). Numerical simulations have enhanced our understanding of these systems primarily by isolating the dominant mechanisms of formation within model sensitivity runs, but also through representing processes that are typically under sampled in the data-sparse regions of polar low formation.

While polar lows appear to have many attributes in common, there is no widely accepted classification scheme for these systems (Rasmussen and Lystad, 1987). Several different schemes have been suggested based on distinctive synoptic setups (Lystad et al., 1986; Rasmussen and Lystad, 1987; Reed, 1987; Businger and Reed, 1989a), and on cloud patterns and shapes observed within satellite imagery (Carleton, 1985; Forbes and Lottes, 1985). Rasmussen (1983) differentiated between the spiral-shaped polar lows and comma-shaped polar lows, and referred to the spiral-shaped case as real or true polar lows. Businger and Reed (1989a) distinguished three types of polar lows based on distinctive synoptic patterns, which provided a distinct degree and distribution of baroclinicity, static stability, and surface latent and sensible heat fluxes. Their three types included (1) the short-wave/jet-streak type (comma clouds) which are characterized by deep, moderate baroclinicity, modest surface fluxes, and a secondary vorticity maximum and PVA aloft; (2) the Arctic-front type which are characterized by shallow baroclinicity, strong surface fluxes and ice boundaries; and (3) the cold-low type which are characterized by weak baroclinicity, strong surface fluxes and deep convection. A combination of these types can exist too. Businger and Reed (1989b) also noted that the comma shape arises as a result of the location of a positive vorticity center within a moderate background wind flow; when the latter is weaker, a spiral pattern tends to develop.

Rasmussen and Turner (2003) tried to improve on the Businger and Reed (1989a) classification scheme by developing a scheme that was based partly on the synoptic setting, and partly on the mechanisms driving polar low formation. Their scheme consisted of seven different types of polar lows observed to occur over the Norwegian Sea (trough systems, cold lows, orographic polar lows, comma clouds, boundary layer fronts, reverse shear systems and baroclinic wave-forward shear types). More recently, a classification scheme for polar low events over the Nordic seas was developed using satellite observations and NCEP re-analysis data that divides polar lows into four types (western polar
lows, eastern polar lows, Greenland lee polar lows and storm track polar lows) (Blechschmidt et al., 2009). These types are distinct in sea level pressure, upper level geopotential height, and the ocean and upper level temperature difference. Pronounced upper level cold troughs or lows of the circumpolar vortex, and large differences between the upper level and sea surface temperatures indicative of strong vertical instability, were common to all of the types.

Debate continues as to whether there are a number of different types of polar lows that are generated by different forcing mechanisms, or whether a range of polar low morphology develops from the same forcing mechanism or combination of forcing mechanisms. A number of mechanisms for the development and intensification of polar lows have been suggested, including baroclinic instability, CISK, air-sea interaction instability/wind-induced surface heat exchange instability, barotropic instability, and the interaction between mobile upper-level PV anomalies and lower-level PV anomalies. While barotropic instability has been observed in association with polar low development (e.g. Mullen, 1979), the contributions from this instability to rapid polar low development appear to be minor (Reed, 1979; Sardie and Warner, 1985) and will not be discussed further here.

Significant baroclinicity has been observed through deep layers of the atmosphere in association with polar lows (e.g. Harrold and Browning, 1969; Mullen, 1979; Reed, 1979; Locatelli et al., 1982). Harrold and Browning (1969) demonstrated that precipitation associated with a polar low passing over England occurred primarily as a result of slantwise ascent, typical of baroclinic disturbances, and Nordeng (1987) found, from their simulations of two polar lows, that the inclusion of a slantwise convection parameterization scheme resulted in the development of stronger low-level winds in both cases. Reed (1979) analyzed two polar low cases in detail and found that they formed poleward of the jet stream where the atmosphere was conditionally unstable and characterized by weak to moderate baroclinicity. A composite analysis of 22 polar lows by Mullen (1979) demonstrated that polar lows developed within deep baroclinic zones on the low-pressure side of well-developed jet streams in regions of strong cyclonic shear. The lower troposphere was conditionally unstable and strongly heated by the ocean in the early stages of development, which was 2-6 K warmer than the air. These conditions were also present in a case study by Rasmussen (1985). Reed and Mullen both concluded that the formation of polar lows is probably a result of baroclinic instability in the presence of low static stability in the lower troposphere as cold air flowing around the large-scale cyclone is heated by the warmer ocean surface. The weak static stability explains the mesoscale size of the system (Staley and Gall, 1977). Mansfield (1974) and Duncan (1977) both showed that polar low development could be supported over realistically short spatial and temporal scales by an environment characterized by a shallow baroclinic layer and low static stability.

Locatelli et al. (1982) found that a number of comma clouds have wind, temperature and precipitation patterns similar to their larger extratropical
cousins that form along the polar front. Relatively intense comma clouds have occasionally been observed to develop over land in the absence of significant surface sensible and latent heat fluxes, thus indicating that baroclinic instability alone may be sufficient to drive these systems (Reed, 1979). Businger (1985) analyzed a composite of 42 cases of well-developed polar lows over the Norwegian and Barents seas and found large-scale conditions similar to those observed previously (Reed, 1979; Mullen, 1979; Rasmussen, 1983). In particular, the lows developed in a strong baroclinic region of very low static stability under a region of positive vorticity advection. His study also showed an outbreak of deep convection at the time of rapid deepening, thus suggesting that latent heating may play a role in the deepening process. The conditionally unstable air mass in which polar lows occur favors the development of convective clouds. A series of modeling studies (e.g. Mansfield, 1974; Duncan, 1977; Staley and Gall, 1977; Orlanski, 1986) demonstrated the importance of reduced static stability and moist processes within a region of baroclinicity in enhancing polar low development. Mudrick (1987) found that the initial stages of polar low development could be explained by dry baroclinic instability. However, the simulated polar lows in the dry baroclinic simulations moved at greater speeds than the observed systems (Reed and Duncan, 1987), thus indicating the potentially important role that latent heat release may have on the structure and growth of these systems. 

Shapiro et al. (1987) were the first to make use of research aircraft data to study polar lows. They suggested that the initiation of polar lows was due to baroclinic forcing on a synoptic scale. It was not possible from their observations, however, to determine the relative roles of mesoscale baroclinicity or convection. Bond and Shapiro (1991) examined two polar lows that developed over the Gulf of Alaska during the OCEAN STORMS field experiment. The polar lows developed near the center of an occluded, synoptic-scale extratropical cyclone, within a zone of low-level mesoscale baroclinic forcing. Satellite and radar imagery demonstrated that convection was insignificant during the growth phase of the polar lows, and they concluded that the polar low development in this case was primarily due to moist, baroclinic processes.

More recently, Yanase and Niino (2005, 2007) performed idealized numerical experiments to examine the impacts of baroclinicity, stratification and average temperature on the development of polar lows. In their first study they found that a polar low develops within a simplified atmosphere due only to the effects of baroclinic instability and diabatic effects. They also observed that a polar low with spiral cloud bands, an “eye” and a warm core develops when the baroclinicity of the basic state is low, whereas a larger-scale polar low with a comma shape cloud developed when the basic state baroclinicity is high. In their second study they showed that baroclinicity is the dominant factor controlling polar low dynamics. When the baroclinicity was weak, positive interactions between the vortex flow and cumulus convection resulted in a
small, weak, almost axisymmetric vortex. Surface friction was found to play an important role in the organization of cumulus convection within the vortex through its impacts on the transport of heat and moisture to the vortex. When the baroclinicity was strong, a larger vortex developed that contained a comma shape cloud pattern. In this case, condensational heating was found to be important, as was the generation of eddy kinetic energy from both the eddy available potential energy and from the mean kinetic energy through the vertical shear. The former case was sensitive to the initial perturbation while the latter was not. Finally, the characteristics of the polar low changed smoothly without any significant regime shifts as the baroclinicity was changed.

Some researchers have found that CISK or other heating mechanisms are important in polar low formation. CISK (Charney and Eliassen, 1964) processes represent a positive feedback between cumulus convection and the low-level convergence involved. In the context of a polar low, convergence associated with the rotating vortex enhances the moisture available for convection, which in turn supplies latent heating that intensifies the circulation, which then provides more moisture to the convection. Numerical simulations suggest that this process appears to play an important role in the development of polar lows (e.g. Økland, 1977, 1987; Rasmussen, 1977, 1979; Bratseth, 1985). Both Bratseth (1985) and Økland (1987) showed that convective heating needs to reach a maximum at lower levels for CISK processes to be significant. Locatelli et al. (1982) analyzed the mesoscale structure of three polar lows and found convective rainbands and cells similar to those that occur with large-scale Pacific extratropical cyclones. The latent heating in the convection may have been the primary mechanism for system deepening as discussed by Rasmussen (1979, 1981).

An observational (Fu et al., 2004) and modeling study (Yanase et al., 2004) was conducted of a 200 km wide polar low that developed over the Japan Sea. The polar low initially developed in association with an E-W orientated cloud band and produced a spiral-shaped cloud pattern with an “eye” structure. A number of model sensitivity tests demonstrated that condensational heating was the primary factor causing the rapid development of this polar low, and that surface fluxes were important for maintaining an environment that supports the development of the low-level vortex. Vortex development was strongly suppressed in the absence of surface fluxes due to the resultant stabilization of the boundary layer.

Miner et al. (2000) recently documented the development of a polar low, similar to the cold-low class described by Businger and Reed (1989a), except that this polar low developed during early autumn over the interior of North America, and subsequently intensified as it moved over the Great Lakes. This cyclone began as a cold-core cyclone but evolved into a warm-core system, and developed an eye and spiral bands. The heat and moisture fluxes from the Great Lakes appear to have played a significant role in the development of the system based on the fact that the cyclone deepened rapidly in the presence of
weak baroclinicity, the surface heat and moisture fluxes were comparable in magnitude to those of other polar low and Category 1 hurricane cases, the low strengthened more at lower levels than at upper levels, and the cyclone evolved from a cold-core structure into a warm-core structure.

Two polar lows developed, one behind the other, on 7 March 2005 over the warm Norwegian current. This area is a region of frequent polar low formation (e.g. Mokhov et al., 2007; Bracegirdle and Gray, 2008). Such serial polar low developments have been observed elsewhere (e.g. Reed and Duncan, 1987; Hewson et al., 2000). Brümer et al. (2009) analyzed the properties of these polar lows which developed within the left exit region of a jet streak. They found that both polar lows had a radius of 100-130 km and extended to about 2.5 km in height. The systems were warm-core systems with temperature anomalies of 1-2 K compared with the ambient environment. In situ mass, water and sensible heat budgets showed that about twice as much of the moisture supply in the subcloud layer came from evaporation compared with convergence, and that, while almost all of the condensed water within these systems was converted into precipitation, only about half of the precipitation at cloud base actually reached the surface. Interactions between the two polar lows observed previously (e.g. Renfrew et al., 1997) could not be detected here.

While some studies of polar low formation have emphasized the role of either baroclinic effects or CISK (or other heating mechanisms), other research has suggested that the interaction of both these mechanisms may be important. Simulations conducted by Sardie and Warner (1983) suggested that both CISK and moist baroclinicity were important in polar low formation, but that moist baroclinic processes alone were insufficient to generate comma clouds. Forbes and Lottes (1985) also observed the importance of both mechanisms from their climatology of Atlantic polar low cases. Sardie and Warner’s (1985) simulation of a Denmark Strait polar low demonstrated that low-level baroclinicity was sufficient to initiate development of the low, but that sensible heating from the surface and latent heating associated with convective and nonconvective heating were essential to sustain the development as the low moved away from the baroclinic zone. In their simulation of a Pacific low, baroclinicity and latent heating were also important, but sensible heating from the surface had little effect on the time scale of the development. They concluded from their modeling studies that the development of polar lows depends on both CISK and moist baroclinicity, but that the relative importance of these processes varied from environment to environment. Moist baroclinicity dominates the formation of the comma-shaped polar lows in the North Pacific, while CISK is the dominant contributor to the spiralform polar low formation in the North Atlantic. The former formed over water, near the polar front and away from ice edges, whereas the latter developed along the marginal ice zone. Businger (1987) showed that ice edge processes were important to the formation of polar lows over the North Pacific. Craig and Cho (1988) also examined the roles of CISK and baroclinic instability by combining the Eady model of baroclinic instability with
wave-CISK. They found that the role of heating was to reduce the static stability, which resulted in faster growth and shorter wavelengths. For small heating rates, perturbations about a baroclinic base state resemble a baroclinic wave. As the heating was increased, the instability transitioned and took on the characteristics of a CISK disturbance.

As only a few polar lows have been observed to occur over land, researchers generally agree that air-sea interaction processes are important to their development. Emanuel (1986) proposed that an air-sea interaction instability (ASII), more recently renamed as wind-induced surface heat exchange instability (WISHE) (Emanuel et al., 1994), plays a significant role in the formation of tropical cyclones. Craig and Gray (1996) provide a detailed comparison of CISK and WISHE. Strong surface winds and decreasing pressure associated with tropical cyclones lead to anomalous sea-surface fluxes of sensible and latent heat, which in turn lead to enhanced temperature anomalies, and hence to decreases in central pressure and increases in surface winds, and so on. Given the large latent and sensible heat fluxes associated with polar lows (e.g. Shapiro et al., 1987), values that are comparable in some cases with those observed in tropical cyclones, this instability may be important in polar low formation. Rasmussen and Lystad (1987) do note, however, that CAPE is large in the cold air outbreaks associated with polar lows and hence the reasoning applied to tropical cyclones may not be applicable to polar lows. Emanuel and Rotunno (1989) used simulations to argue for ASII and against CISK as the dominant energy source in polar low development. They did, however, note that a triggering mechanism is needed before the ASII occurs.

Craig and Gray (1996) used a cloud-resolving axisymmetric simulation to examine the relative contributions of CISK and WISHE to polar low intensification. As they point out, for both WISHE and CISK, the rate of growth is controlled by boundary layer processes. In the WISHE theory, the rate limiting process is the rate at which latent and sensible heat is fluxed from the ocean surface. For the CISK mechanism, rapidly acting surface fluxes are assumed to maintain CAPE, and hence the rate limiting process in this theory is the fractional mass convergence. Thus, the rate of growth is controlled by heat and moisture fluxes in WISHE, but by frictional convergence in CISK. Craig and Gray’s modeling results demonstrate that the rate at which their simulated cyclone systems intensified, increased with increasing values of the heat and moisture transfer coefficients, and hence with increasing surface heat and moisture fluxes. Frictional convergence was found to be of secondary importance. Hence they conclude from their results that the intensification of modeled polar lows is due to WISHE, rather than CISK.

More recently, the role of upper-level PV anomalies in the initiation and development of polar lows have been recognized (e.g. Mullen, 1983; Businger, 1987; Nordeng, 1990; Businger and Baik, 1991; Shapiro et al., 1987; Montgomery and Farrell, 1991, 1992; Nordeng and Rasmussen, 1992). Nordeng (1990) suggested that polar low development occurs in two phases in which
baroclinic interaction first occurs between an upper-level trough and a surface
disturbance, followed by an ASII. The results of simulations by Nordeng and
Rasmussen (1992) indicated that upper-tropospheric forcing plays an important
role in the organization of the ascent that leads to the spinup of the lower
tropospheric vortex. Through the use of a two-dimensional semigeostrophic
Eady model, Montgomery and Farrell (1991) found that an initial disturbance
with interior PV exhibited strong baroclinic coupling between the upper and
lower disturbances compared with those situations of uniform interior PV.
Their simulations also demonstrated a two phase development of polar lows,
beginning with an initial baroclinic growth phase, and being followed by a
long, slow intensification due to diabatic effects. Montgomery and Farrell
(1992) then used their results from a three-dimensional nonlinear geostrophic
momentum model to further develop their two stage conceptual model of polar
low development. During the first stage of development, a stage they refer to
as induced self-development, an interaction occurs between upper-level and
lower-level PV anomalies in a nearly moist neutral baroclinic atmosphere.
Ascending motion ahead of the advancing trough results in rapid spinup in
the lower levels and the production of PV anomalies at low and midlevels,
which enhances the baroclinic interaction between the upper-level and lower-
level systems. In the second stage, referred to as diabatic destabilization, a
secondary intensification occurs as a result of the production of low-level PV
in ascending regions through diabatic processes. The secondary phase appears
to progress more slowly than the primary phase. Diabatic destabilization thus
provides a mechanism through which polar lows can maintain or enhance their
intensity until they reach land, as is often observed. While CISK and/or ASII
may contribute to system enhancement during the later stage of development,
neither appeared to be necessary.

Douglas et al. (1995) compared the characteristics of a 300 km diameter
polar low that developed over the northern Gulf of Alaska during the Alaska
Storms Program (Douglas et al., 1991) and a 400 km diameter polar low
that formed along the ice edge of east Greenland during the 1989 CEAREX
experiment. Frontal zones were observed to occur with the CEAREX polar
low, and the cloud field was more similar to that of an extratropical cyclone
than the Alaskan polar low. The vortex of the Alaskan low was warmer than
its surroundings at lower levels, and suppressed cloudiness characterized the
vortex center, possibly occurring as a result of warm air seclusion similar to
in the Shapiro-Keyser model of extratropical cyclone development discussed
previously in this chapter. Unlike the Alaskan low, the CEAREX low did not
have a clearly defined warm inner core. Convection was deeper in the Alaskan
case. In spite of the fact the CEAREX polar low occurred over open water,
significant intensification did not appear to occur, suggesting that surface heat
and moisture fluxes were not necessary for further development, although it is
possible that the conditions were not suitable for ASII processes to occur. Also,
as the convection was shallow, this case suggests that the upper-tropospheric
short-wave was not coupled to the lower-level vortex via convection. Instead, it appears that the formation and intensification of the polar low occurred through the interaction between an upper-level mobile PV anomaly and a lower-level PV anomaly (Montgomery and Farrell, 1992).

Finally, Bresch et al. (1997) simulated the development of a polar low that occurred over the western Bering Sea in order to examine the processes important to its development. Observations of this case study showed that the polar low formed in a region of moderate low-level baroclinicity near the ice edge when a region of high PV associated with an upper-level trough advected into the region. Various sensitivity tests were conducted including, turning off the surface fluxes which failed to develop a polar low, switching on the surface fluxes after 24 hours in which only a weak low developed, and increasing the distance from the ice edge which had small positive effects. Experiments which included surface fluxes but no latent heating, and then latent heating but no surface fluxes, produced polar lows of similar intensity, demonstrating that these two processes are equally effective in enhancing development. They concluded from their experiments that the development of polar lows of this type are similar to that of marine extratropical cyclones, and that polar lows require the interaction between a mobile upper-level PV anomaly and a low-level PV anomaly generated either through thermal advection or diabatic heating.

From the discussion above it seems that neither a single mechanism nor a single structure or shape appears to govern the initiation and development of polar lows, and that the mechanisms are not discrete but rather continuous. Perhaps as Bresch et al. (1997) suggest, polar lows should be viewed as part of a wide spectrum of maritime cyclones that varies based on the strength of the upper-level forcing, tropospheric stability, degree of baroclinicity, deep moist convection, the amount of latent heat release and the magnitude of the surface heat and moisture fluxes, rather than as a discrete type of oceanic cyclone.

### 10.3.7. Lake-Effect Storms

During the fall and winter months, when cold Arctic air associated with the passage of extratropical cyclones sweeps across the relatively warm waters of the Great Lakes, or other large lakes, local, often heavy snowstorms occur along the lee shores. Most of the heavy snowfall does not occur during the passage of cold fronts, but several hours afterward. In some of the more severe storms, snowfall accumulations of more than 75 cm per day are not uncommon (Wiggin, 1950) and snowfall rates of 30 cm hr\(^{-1}\) have been reported (E.S.S.A., 1966). A single storm event produced 175 cm (68.9 in) of snow in Ohio (Schmidlin and Kosarik, 1999), and totals of 150-250 cm over several days have been observed (Niziol, 1989). An important feature of these storms is their persistence for several days over limited and sharply defined regions. These events often also pose a forecasting hazard as one location may receive more than 100 cm of snow over several days while locations only 20 km may receive just a trace.
Lake effect events have previously been defined as those events where heat and moisture fluxes from a lake surface result in the development of an internal convective boundary layer and associated clouds (e.g. Lenschow, 1973; Chang and Braham, 1991; Kristovich and Laird, 1998). An excellent overview of the climatology, characteristics and factors influencing lake effect snow is provided by Niziol et al. (1995), a summary of which is included here. Lake effect precipitation tends to be greatest earlier in the cold season, as the decline in the lake temperatures and the air-lake temperature difference, and the increase in lake ice, all act to reduce fluxes of heat and moisture from the surface of the lake as winter progresses (Jiusto et al., 1970). In general, lake effect processes are most effective when the prevailing wind blows across the greatest fetch of water. Orography on the lee side of the lake may further enhance lake effect snow. Muller (1966) observed an increase of 12-20 cm in the annual snowfall for every 330 m increase in elevation to the lee of the lakes. The transfer of sensible and latent heat from the lake to the atmosphere can trigger convection that is typically organized into long, linear features known as cloud streets or cloud bands, the orientation of which is determined by the wind profile between the lake surface and the subsidence inversion associated with the Arctic air mass. The bands tend to be aligned parallel to the steering wind, the exception being when thermally-induced circulations dominate in conditions of weak winds. While lake effect snow is generally thought to occur on the eastern sides of the Great Lakes in association with typical westerly flow over the lakes, cyclones that track to the south of the Great Lakes can produce lake effect snow on the western lake shore due to the easterly flow over the lakes. Lake effect processes and frontal processes can also interact, in that the mid- to upper-level frontal clouds can seed the lower lake-induced clouds (Kristovich et al., 2000). Lake effect snowstorms appear to be produced by complex interactions between processes on a wide range of scales from the microphysical through to meso-α scale, and it is this complexity and range of spatial and temporal scales of the processes that make lake effect snow difficult to forecast. Several recent studies have also shown that lake effect snowfall has increased over the twentieth century, a statistic attributed to warmer lake waters and decreased ice cover (Norton and Bolsenga, 1993; Burnett et al., 2003; Kunkel et al., 2009).

Arctic air masses are often accompanied by a strong subsidence inversion, which limits the depth to which convection can grow. The fluxes of heat and moisture from the lake surface can lift and/or erode the inversion (Lavoie, 1972). Thus the low-level instability (air-lake temperature difference) and the depth of the unstable layer (height and intensity of the inversion) are important factors controlling the intensity of lake effect snow. Also, differences in the temperature and surface roughness between the lake and the shore generate low-level thermal gradients and frictional convergence, which result in organized vertical motion that can aid in lifting the capping inversion, thereby enhancing convective development and precipitation (Hjelmfelt, 1990; Niziol et al., 1995). A conceptual model of a major snowstorm over Lake Erie developed by
PART II  The Dynamics of Clouds

FIGURE 10.26  Schematic cross section of lake effect snow. (After Davis et al. (1968))

Davis et al. (1968) is shown in Fig. 10.26. Illustrated is a stratocumulus cloud that deepens over the lake as it nears the lee shore. Corresponding to the deepening cloud layer is an upward displacement of the capping inversion. The faster falling, heavily rimed snow crystals such as graupel particles precipitate just onshore. This is also the location of the heaviest snowfall amounts. Lightly rimed dendrites and plates are carried further inland. Aggregates of snow crystals having terminal velocities between graupel and unrimed single crystals settle somewhere in between.

Four general types of lake effect storms have been identified over Lake Michigan (Braham and Kelly, 1982; Forbes and Merritt, 1984; Hjelmfelt, 1988). Hjelmfelt (1990) was able to simulate all four types simply by varying the lake-land temperature difference, the wind direction and strength, and the static stability. Schematics of these four storm types are shown in Fig. 10.27, and include the following:

(1) Widespread coverage/wind parallel bands (e.g. Kelly, 1982, 1984; Braham, 1986; Kristovich, 1993; Steve, 1996): the winds in association with these bands show no reversal on the lee shore; the bands may become organized into wind parallel bands (Kelly, 1982, 1984), open convective cells (Braham, 1986) or a combination of these two types (e.g. Kristovich et al., 1999; Cooper et al., 2000); they occur during strong westerly winds, strong static stability, and strong lake-land temperature differences. With band widths of 2-4 km and spacings of 8-20 km, these bands tend to produce widespread snowfall of low intensity. Wind parallel bands may join to form less regular cloud morphologies.

(2) Shoreline parallel bands (e.g. Braham, 1983; Hjelmfelt and Braham, 1983; Schoenberger, 1986): these bands occur with a well-developed land breeze on the lee shore (Braham, 1983); they occur during moderate westerly winds, weaker static stability and a strong lake-land temperature difference.

(3) Midlake bands: these bands are associated with low-level convergence centered over the lake (e.g. Passarelli and Braham, 1981); they occur when
westerly and easterly winds from both sides of the lake converge, and in association with moderate lake-land temperature differences and weak static stability.

(4) Mesoscale vortices: these vortices are associated with a well-developed cyclonic flow in the boundary layer, usually occurring with weak surface-pressure gradients and a ridge of high pressure centered over the lake or to the west of the region (e.g. Peace and Sykes, 1966; Forbes and Merritt, 1984; Pease et al., 1988); they occur under conditions of strong lake-land temperature differences, weak northerly winds and weak stability.

A mix of these morphologies may also occur (Shoenberger, 1986). It should be noted that the emphasis in the modeling study by Hjelmfelt (1990) was on the east coast of Lake Michigan, and hence the conditions listed above are somewhat specific to this region. These morphologies could however be expected to exist under other flow regimes over different lakes. Five snow band types have been identified for the eastern Great Lakes (Niziol et al., 1995). In the classification above the wind parallel bands were not separated into those that form when the winds blow parallel to the long axis of the lake and those that form when they blow perpendicular to the long axis of the lake, an effect that turns out to being important for the eastern Great Lakes.

A number of typical lake effect morphologies occurring simultaneously are evident in the GOES visible satellite image shown in Fig. 10.28 (Laird et al., 2003a). A widespread coverage event, consisting of both wind parallel bands and cellular convection is evident over and downwind of Lake Superior, being associated with unidirectional winds across the lake, and producing light to moderate snowfall over and downwind of the lake. A shoreline band is evident over Lake Michigan. This band develops in association with a thermally-driven land breeze circulation and/or prevailing winds nearly parallel to the major axis of the lake, and typically produces heavy snow and the strongest updrafts of all four band types. Finally, a vortex is evident over Lake Huron. These events

![Figure 10.27](image-url)
are typically associated with light snow and weak land breezes, and may have a cloud-free region in the center.

The most common convective pattern over most lakes appears to be the widespread type, which accounts for nearly 60% of cloudiness during December and January (Kelly, 1986; Kristovich and Steve, III, 1995). Also, the widespread wind-parallel bands of convection associated with boundary layer rolls are generally associated with strong cross-lake winds. Convective bands that develop parallel to the long axis of a Great Lake tend to occur in conditions of high air-lake temperature differences and weak contributions from the cross-lake wind component. The midlake bands and the shoreline bands tend to be wider than the wind parallel bands (Braham and Dungey, 1995), and can produce very heavy snowfall over and near the lake shore. Midlake bands and shoreline bands both form from a combination of dynamic and diabatic forcing. Midlake bands may develop when strong winds blow along the lake major axis resulting in enhanced surface heat fluxes, the generation of a mesoscale low pressure over the lake, the development of a land-breeze circulation, and the subsequent development of a midlake convergence zone and associated midlake snowband (e.g. Braham and Kelly, 1982; Niziol et al., 1995). Shoreline parallel bands can develop when the winds are calm, and the strong horizontal temperature gradients associated with large lake-land temperature difference, results in the development of a mesoscale overlake low, the formation of a land breeze, and the development of the shoreline parallel bands (e.g. Passarelli and Braham, 1981). The shoreline bands may even occur in the absence of a land breeze, providing strong convergence develops on the lee side of the lake, although these systems will tend to be weaker and shorter-lived than those
associated with a well-developed land breeze. Given the similarities in their formation, midlake and shoreline bands are occasionally grouped together into the same category (e.g. Laird et al., 2003a,b). Lake vortices appear to be a weak version of the polar lows described above. They have been observed to occur with shoreline parallel or midlake snowbands (e.g. Pease et al., 1988; Grim et al., 2004), or with other lake vortices (Laird, 1999). These systems range in size from 10 km (Schoenberger, 1986) to over 100 km (Forbes and Merritt, 1984). In the vortex studies of Schoenberger (1986) and Grim et al. (2004), a land breeze front propagated westward into an area of horizontal convective rolls. Upward tilting of the rolls by the land breeze front appeared to be responsible for the vortex generation, although the observations in the Grim et al. (2004) could not confirm this.

The most intense snowstorms are the shoreline parallel bands which occur over the Great Lakes following the passage of a cold front when the following conditions are met (Sheridan, 1941; Wiggin, 1950; Rothrock, 1969):

1. the difference between the lake surface temperature and the 850 mb temperature exceeds 13 °C corresponding approximately to a dry adiabatic lapse rate between the surface and 850 mb;
2. an onshore wind with an overwater fetch of more than 100 km is present;
3. over water, low-level wind speeds are moderate to strong; and
4. the height of the boundary layer capping inversion exceeds 1000 m.

Ideas regarding lake effect snow were developed as early as the 1950s using observations. Wiggin (1950) noticed that the width of the snowbands was proportional to the depth of the cold air, while Peace and Sykes (1966) suggested that pressure troughs and narrow zones of confluence were important to snowband formation. A number of numerical simulations were also conducted of lake effect storms in the 1970s and 1980s. Using a simple mixed-layer model, Lavoie (1972) demonstrated the importance of wind speed and direction to the strength of the mesoscale disturbance, with the longest fetch of air across the lake surface producing the strongest disturbance. Both Lavoie (1972) and Ellenton and Danard (1979) showed that the heat and moisture fluxes from the lake were the principal causes of precipitation, while shoreline convergence and orographic lift played secondary roles. Likewise, Hsu (1987) showed that the coupling between surface heat fluxes and winds were responsible for the formation of patterns of low-level convergence that generated precipitation. Hjelmfelt and Braham (1983) simulated the lee-shore parallel band over Lake Michigan, including the formation of a surface mesolow beneath the major cloud band. Latent heat release was found to be important in strengthening the convection, although the land-breeze circulation that was associated with the band formed even without latent heat release. Ballentine et al. (1992), like Hjelmfelt (1990), made use of sensitivity tests to examine the significance of water temperature, land and air temperature, wind speed and direction, and the humidity on a lake effect snowband.
Recent advances in both satellite and radar technology, as well as computer modeling, have enhanced our understanding of lake effect processes. Agee and Gilbert (1989) made use of data from the Lake-Effect Snow Studies (LESS) field campaign and identified a penetrative convective layer located between the top of the fully mixed layer and the part of the inversion layer that has not been modified. This penetrative convective layer creates a large amount of turbulent exchange between the mixing layer and the capping inversion. Chang and Braham (1991) examined the development of the convective thermal internal boundary layer over Lake Michigan and found that the average slope of the boundary layer was 1% over a fetch of ∼125 km. Byrd et al. (1991) used sounding data to investigate lake effect bands and suggested that the depth of the mixed layer may be more important than the strength of the instability precipitation intensity, while Burrows (1991) identified low-level mass divergence as the primary parameter controlling the amount of snow produced. Recent observations from the Lake-Induced Convection Experiment (Lake-ICE) demonstrated the importance of air-lake fluxes and the vertical structure of the atmosphere in the development of lake-effect storms (Scott and Sousounis, 2001).

The presence of ice on a lake surface reduces the heat and moisture fluxes from the surface, and thus will influence lake effect storms (e.g. Niziol et al., 1995; Laird and Kristovich, 2004). Using observations, Gerbush et al. (2008) estimated that surface sensible heat fluxes increased to open water values as the ice concentration over Lake Erie decreased from 100% to 70%, while numerical simulations by Zulauf and Krueger (2003) demonstrated that increases in the ice thickness from 0 to 10 cm (10 to 20 cm) were accompanied by decreases in the sensible heat flux of ∼50% (25%). The significance of thinner ice on the enhancement of snowbands has also been observed recently over Lake Erie (Cordeira and Laird, 2008). The lake temperature also appears to influence the lightning characteristics of lake effect storms, with lightning occurring predominantly between September and December due to the warmer lake temperatures and the greater convective cloud depths (Moore and Orville, 1990). Lake effect storms with lightning also have significantly higher temperatures and dewpoint temperatures in the lower troposphere, lower lifted indices, higher lake-induced equilibrium levels and CAPE, lower wind shear, and an increase in the mean height of the −10 °C level compared with non-electrified storms (Schultz, 1999; Steiger et al., 2009). Lightning producing storms were also strong, single-band storms.

Several studies of lake effect events have been more climatological in nature. Ellis and Leathers (1996) showed for the eastern Great Lakes that, while the large-scale synoptic situation was similar for each lake effect type, variations in the sea level pressure patterns, 850 mb temperatures and heights, 500 mb heights, seasonality, fetch and the strength of the flow produced significant differences in the location, magnitude and frequency of the lake effect snowfall. In their study of autumnal (Sep-Nov) lake effect precipitation downwind of Lake
Erie, Miner and Fritsch (1997) found that lake effect precipitation occurs on approximately one out of every five days, with a diurnal peak in precipitation intensity during the afternoon and evening. They also noted that the greatest number of lake effect days occurred in October, that lake enhanced precipitation actually begins in late summer (when the lake temperature already starts to exceed that of the surrounding air), and that the precipitation transitions from rain to snow during November. Kristovich and Spinar (2005) examined the diurnal characteristics of lake effect events over Lakes Superior and Michigan and found a morning maximum and an afternoon/evening minimum in the frequency of lake effect precipitation.

The impacts of the intensity, track and duration of synoptic-scale systems were found to have a significant effect on lake effect storms (Ballentine et al., 1998; Schmidlin and Kosarik, 1999). Lake effect processes that are enhanced during the passage of a synoptic low pressure system have been referred to as lake-enhanced snowfall (Eichenlaub, 1979). More recently, Lackmann (2001) observed from a composite of 32 lake effect events over Rochester, New York that all were accompanied either by a mobile upper level trough or a closed low at the 500 mb level. An examination of a particularly intense event suggested that the upper trough results in an increase in the inversion altitude and relative humidity in the lower troposphere. Liu and Moore (2004) developed a climatology of lake effect snowstorms over southern Canada for the years 1992-1999. They observed that a low pressure and cold-temperature anomaly over Hudson Bay, north of the Great Lakes, provides an environment conducive to lake effect storms over southern Ontario, and that the track of the low can have a significant impact on the development of these storms.

Studies of lake effect snowstorms have also been performed for smaller lakes including the Great Salt Lake (e.g. Carpenter, 1993; Slemmer, 1998; Steenburgh et al., 2000; Steenburgh and Onton, 2001; Onton and Steenburgh, 2001), Lake Champlain (e.g. Tardy, 2000; Payer et al., 2007; Laird et al., 2009a), and the New York State Finger Lakes (e.g. Cosgrove et al., 1996; Watson et al., 1998; Sobash et al., 2005; Laird et al., 2009b), as well as other small lakes in the United States (e.g. Wilken, 1997; Cairns et al., 2001; Schultz et al., 2004). Some of the lake effect events have produced significant amounts of snow in spite of the lake size and the localized nature of the storms. Steenburgh and Onton (2001) examined the development lake effect snowstorms over the Great Salt Lake and found that the primary snowband formed along a land-breeze front. They concluded from their study that even though the Great Salt Lake is relatively small in comparison with the Great Lakes, that thermally-driven circulations and banded precipitation structures similar to those over the Great Lakes can still occur. In a follow-on paper, Onton and Steenburgh (2001) conducted a modeling study of snowband development over the Great Salt Lake. They found that moisture fluxes from the lake were necessary for the development of the snowband. However, the saline composition of the lake
resulted in a reduction in the moisture fluxes compared to a freshwater lake, causing a 17% decrease in snowfall.

Widespread lake effect events are often composed of boundary layer rolls. Roll vortices have been observed to occur when the convective pattern is organized into wind-parallel bands of heavier precipitation. These rolls have been shown to influence the surface layer fluxes of heat and moisture (e.g. LeMone, 1976). Numerous factors have been suggested for their formation including high wind shear, curvature of wind speed profiles parallel to the roll axes, inflection points in the wind profiles, boundary layer mean shear, and gravity waves (Kristovich et al., 1999 and the references therein). However, Kristovich (1993) found that none of these criteria were met in every one of the cases of roll convection that he studied, and suggested that low-level shear was the primary factor in roll development. Transitions between nonroll and roll convection have been observed. Such transitions may be associated with large changes in the snow spatial coverage and mean snowfall rate (Atkinson and Zhang, 1996). The two most important factors for differentiating between roll and cellular convection according to Atkinson and Zhang (1996) are atmospheric dynamic forcing and thermal instability.

Kristovich et al. (1999) examined the mechanisms that result in the transitions between boundary layer rolls and the more cellular convective structures observed in association with lake effect snow. They found that roll formation occurred following increases in the low-level wind speed and speed shear below \( \sim 0.3z_i \) where \( z_i \) is the boundary layer depth. Mass overturning rates were greatest at midlevels in the boundary layer when rolls were dominant and decreased when cellular-type convection was dominant. Mean snowfall rates demonstrated little change with the transitions from one form to the other, but the heaviest snow was more concentrated in the updraft regions when the rolls were dominant. Cooper et al. (2000) found from their simulations that the variation in wind speed shear below 200 m played a major role in the degree of linearity of the convection; directional shear was not a requirement for roll convection. Tripoli’s (2005) idealized simulations showed that shoreline geometry effects were sufficient to generate a near-surface streamwise vorticity, which served as the seed for roll development at the most efficient mode of roll convection.

Combined heating and moistening from all of the Great Lakes has been found to influence the atmospheric conditions near the individual lakes, which in turn affects the lake effect storms that may develop. This combined influence of lakes is referred to as the lake-aggregate effect (e.g. Sousounis and Mann, 2000) or collective lake disturbances (Weiss and Sousounis, 1999). Lake-aggregate circulations are on the meso-\( \alpha \) scale (200-2000 km) compared with the meso-\( \beta \) scale (20-200 km) of the individual lake circulations. An upstream lake can impact the distribution of local lake effect snow (Byrd et al., 1995), and the aggregate effect has been found to enhance lake effect precipitation in northern lower Michigan and in southern Ontario, but reduce it in regions south and east
of Lakes Erie and Ontario (Sousounis and Mann, 2000). A factor separation analysis (Stein and Alpert, 1993) of model output to assess the contributions from adjacent lakes (both upstream and downstream) to lake-effect storm evolution showed that interactions amongst lake-scale processes contributed to the development of the regional-scale disturbance and that the relative influence of the adjacent lakes increased as the collective lake disturbance matured (Mann et al., 2002). During the development of the collective lake disturbance, contributions from the lake-lake interactions tended to offset the individual lake contributions, however, as the regional-scale disturbance matured, the lake-lake interactions then enhanced the individual lake contributions. For example, the eastern Great Lakes (downstream lake) were found to reduce the precipitation amounts over southern Lake Michigan by as much as 25% through reduced convective instability effects, and the impacts of Lake Superior (upstream lake) on Lake Michigan caused a delay in the maximum snowfall occurrence. Finally, as the collective lake disturbance matured, the Lake Superior processes were found to have a significant effect on the snowband morphology over Lake Michigan.

Laird et al. (2003a) conducted numerous simulations to examine the factors controlling the meso-β scale circulations of lake effect storms. They found that the morphological regimes could be predicted using the ratio of wind speed to maximum fetch distance (U/L). Environmental conditions with low values of U/L (~< 0.02 m s⁻¹ km⁻¹) were associated with vortex circulations, those with moderate values of U/L (between 0.02 and 0.09 m s⁻¹ km⁻¹) with shoreline bands, and those with higher U/L values (>∼ 0.9 m s⁻¹ km⁻¹) with widespread events. They also noted that transitions from one regime to the next were continuous, and that within transitional regions, more than one regime may exist. Laird et al. (2003b) noticed a change in morphology from vortices to bands to widespread coverage as U/L increased, and that the conditions supporting multiple morphologies were more favorable for elliptical lakes than circular lakes. Laird and Kristovich (2004) then assessed the U/L criteria using observational data and found that, even though the U/L provides useful information regarding the band morphology, the criterion only provides limited use when used for actual forecasting of the band morphology.

Many of the more recent studies of lake effect bands and precipitation have been associated with the Lake-Induced Convection Experiment (Lake-ICE) and the Snowband Dynamics Project (SNOWBAND) conducted during the winter of 1997/98 (Kristovich et al., 2000). Kristovich et al. (2003) examined the microphysical and thermodynamic characteristics of the boundary layer and convective organization of a lake effect event that occurred over Lake Michigan during this campaign and found that, while the horizontal scale of the convective structures grew across the lake, it did so less rapidly than the depth of the convective boundary layer, thus the aspect ratio decreased, which is contrary to previous findings. Miles and Verlinde (2005a,b) observed transient linear organization of the convection for a case during Lake-ICE,
even though the conditions should only have supported cellular organization. Mode switching has been previously observed (e.g. Braham, 1986; Kristovich et al., 1999). Miles and Verlinde (2005a) found that the transition between the linear and cellular modes showed no correlation with the mean or low-level shear, surface buoyancy fluxes or stability parameters, thus suggesting that those factors that normally control the linear organization of convection did not affect the transition. In the second paper in this series, Miles and Verlinde (2005b) investigated the possible role of non-linear interactions between different scales of motion in the transient linear organization and found that the net nonlinear interactions between the roll and turbulence scales were significant in magnitude. Nonlinear interactions may thus help to explain the observed transitional linear organization.

Finally, interactions may occur between the synoptic-scale circulations of extratropical cyclones and mesoscale circulations of lake effect storms through boundary layer growth rates and seeding effects (e.g. Lenschow, 1973; Agee and Gilbert, 1989; Chang and Braham, 1991). Schroeder et al. (2006) examined the interaction of a synoptic cyclone with the convective boundary layer in a lake effect event over Lake Michigan. They found that both the precipitation rates and the growth of the convective boundary layer were enhanced through seeding from the clouds above, and the impact of the cyclone on the thermodynamic characteristics of the air over the lake. Not only was the convective boundary layer growth enhanced due to the interaction of the convection with a layer of reduced stability above, but the convective boundary layer was also deeper in the seeded regions compared with the non-seeded regions and appears to be related either to enhanced latent heat release or to mesoscale updrafts. The snowfall rates were similar to those produced in previous lake effect events when the surface heat fluxes were much larger, but the system was not interacting with an extratropical cyclone.

In summary, three major streams of air are involved in the three-dimensional circulation of extratropical cyclones. Because of their different origins and the variety of dynamical and physical processes that they experience in their passage through the cyclone system, the characteristics of the environment vary greatly with time and space. These environments, with different temperatures, moisture content, vertical motion, static stability, and wind shear, host the enormous variety of mesoscale phenomena and clouds present in extratropical cyclones.

10.4. MIDDLE- AND HIGH-LEVEL CLOUDS

10.4.1. Introduction

Middle- and high-level clouds play an important role in both the radiative and water budgets of the earth, an effect made all the more important given their extensive coverage. On an annual average, clouds cover between 55% and 60% of the earth (Matveev, 1984), and much of this cloud cover
FIGURE 10.29 2.5 × 2.5° average of cloud type distributions based on CloudSat measurements over the initial 1-year period. (From Sassen and Wang, 2008)

consists of vast sheets of middle (altostratus and altocumulus) and high (cirrus, cirrostratus, and cirrocumulus) clouds. Global distributions of cloud type generated using CloudSat data (Sassen and Wang, 2008) are shown in Fig. 10.29, and are compared with those from the International Satellite Cloud Climatology Program (ISCCP) (Rossow and Schiffer, 1999) and surface observation reports (Hahn and Warren, 1999) in Table 10.2. As is evident from the “all cloud” fraction panel in Fig. 10.29a, cloud fractions over the mid-latitude storm tracks are high in both the northern and southern hemispheres, reaching a value of ~80% in the southern storm track. High cloud distributions (Fig. 10.29b) are greatest over the ITCZ, coinciding with deep convective systems (Fig. 10.29h; Mace et al., 2006a). The uneven distribution of high clouds over the mid-latitudes is due to the variety of cirrus-forming mechanisms, including those associated with the synoptic jet stream, deep convection and orographic uplift. High frequencies of middle clouds are found poleward of 30° (Fig. 10.29c,d). The distribution of altostratus corresponds to the extratropical cyclone storm tracks, whereas altocumulus clouds are more closely connected with deep convection. Middle and high clouds can produce significant precipitation in association with organized tropical and extratropical cyclonic storm systems.
TABLE 10.2 Comparison of 1-year CloudSat global cloud type frequency averages over land and ocean with annual means of extended surface observer reports (Hahn and Warren, 1999) and ISSCP annual means from 1986-1993 (Rossow and Schiffer, 1999).

<table>
<thead>
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<th>Type</th>
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<th>CloudSat Ocean</th>
<th>Surface Land</th>
<th>Surface Ocean</th>
<th>ISCCP Land</th>
<th>ISCCP Ocean</th>
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<tr>
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<td>17.2</td>
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</tr>
<tr>
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<td>10.7</td>
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<td>3.2</td>
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<td>2.4</td>
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</tbody>
</table>

Source: From Sassen and Wang (2008)

Middle-level clouds refer to altocumulus and altostratus, which may be composed entirely of liquid water or a mix of liquid water and ice. Nimbostratus may be classified either as middle or low clouds and are not considered here. While the elevation of middle-level clouds may vary considerably with season and latitude, a typical elevation in middle latitudes is \( \sim 3 \) km, or about 700 mb. High-level clouds refer to cirrus (including cirrostratus and cirrocumulus) clouds which are composed primarily of ice, although cirrocumulus may be mixed-phase clouds. The elevation of cirrus clouds may also vary considerably; a typical height is 10 km, or 250 mb. Cirrus trap longwave radiation and reflect shortwave radiation, the net effect of which is dependent on cloud thickness; precipitating ice crystals from cirrus clouds can trigger glaciation of warm clouds such as altocumulus clouds, thereby changing the precipitation efficiency of these clouds; the scavenging of aerosols and soluble trace gases and the subsequent ice crystal sedimentation can lead to their redistribution throughout the troposphere; and optically thin cirrus in the tropical tropopause layer (TTL) can dehydrate air entering the tropical tropopause layer (e.g. Holton and Gettleman, 2001). Thus these clouds play an important role in the water budget and the radiation balance in the upper troposphere, and hence in the global climate.

The physical properties of middle- and high-level clouds and their association with extratropical weather systems are discussed in this section, as are the factors influencing their development, and aspects that need to be considered when representing such clouds in GCMs. Given the location of extratropical weather systems, the focus of this chapter will be on high- and middle-level clouds in the mid-latitudes, although comparisons with their tropical counterparts will be made where applicable. Arctic clouds will not be considered here.
10.4.2. Characteristics and Formation of Middle- and High-Level Clouds

Middle- and high-level clouds form when the lifting of moist air, often on large scales as in association with extratropical cyclones, adiabatically cools the air to its dew-point temperature. Even at temperatures well below freezing, the clouds almost always form first as water droplets, with the initial condensation occurring on the typically abundant CCN. Direct transformation from water vapor to ice crystals (sublimation) is possible, but rarely occurs due to the scarcity of sublimation nuclei in the atmosphere. The evolution of the cloud after the formation of the first cloud drops depends on two factors - the temperature of the cloud and the vertical motion. At temperatures below freezing, ice processes are likely to be important because of the relative abundance of freezing nuclei in the atmosphere. These nuclei trigger the freezing of supercooled water to ice. Once temperatures drop below $\sim-33 \, ^\circ C$, homogeneous freezing of supercooled droplets becomes likely, and at temperatures below $-40 \, ^\circ C$ and relative humidities greater than $\sim 80\%$, homogeneous freezing of haze droplets becomes likely. Once ice appears, the difference in saturation vapor pressure over ice and water favors the rapid growth of ice crystals at the expense of the water drops, referred to as the Bergeron-Findeisen process.

If the upward vertical velocities are sufficiently strong ($10 \, \text{cm} \, \text{s}^{-1}$ or greater) and persist for enough time to cause further cooling of the layer, the water drops or ice crystals will grow to sizes wherein fall velocities relative to the air motion become significant and precipitation will occur. The physical processes that produce this growth are collision and coalescence in all water clouds. The Bergeron-Findeisen process, riming of supercooled cloud droplets and aggregation among ice crystals are important in clouds containing both ice and water. For weaker upward velocities ($1-2 \, \text{cm} \, \text{s}^{-1}$), the temperature change resulting from adiabatic expansion is reduced and the temperature changes from radiative effects become more significant. Cooling at cloud top due to longwave radiative flux divergence and warming at cloud base as a result of the longwave radiative flux convergence destabilize the cloud layer and lead to internal convective circulations in the cloud layer. This destabilization is aided by the release of latent heat near cloud base and evaporative cooling near cloud top.

Middle- and high-level clouds frequently occur in layers or sheets of great horizontal dimensions (hundreds or even thousands of kilometers). Because their horizontal dimensions are much greater than their vertical extent, which is typically a kilometer, they are called stratiform or layered clouds. Although this terminology suggests a statically stable lapse rate, thin layers of conditional or convective instability are often present, and these instabilities are conducive to the formation of small-scale convection embedded in the cloud layer. The type of middle or high cloud is strongly affected by the static stability within the cloud layer. Altostratus clouds are produced by the lifting of a layer of air in which the lapse rate is less than the saturated adiabatic lapse rate. For
layers in which the lapse rate exceeds the saturated adiabatic lapse rate, vertical convection results in altocumulus clouds. For high clouds, absolutely stable layers are associated with cirrostratus clouds, while cirrus and cirrocumulus clouds occur when the lapse rate exceeds the saturated adiabatic lapse rate. Alternating regions of upward- and downward-moving air associated with gravity waves in a stable cloud layer can result in a banded cloud structure.

10.4.2.1. Middle-Level Clouds

Altocumulus and altostratus clouds together cover about 22% of the earth’s surface (Warren, 1988). They often extend through the freezing level and thus may either be liquid-phase or mixed-phase clouds. The relative proportions of liquid water and ice have important implications for microphysical processes, aircraft icing, radiative transfer, and remote sensing. However, thin midlevel clouds are not well represented within GCMs. In a recent study of GCMs, all of the models considered, significantly under-predicted thin, midlevel clouds such as altocumuli, and over-predicted thick midlevel clouds such as nimbostratus (Zhang et al., 2005). Others have also shown that thicker midlevel clouds are also not well represented in large-scale models (e.g. Ryan et al., 2000; Xu, 2005). Compared with other cloud systems, relatively few studies have been conducted on the structure and characteristics of mid-level, mixed-phase clouds. The majority of these studies (e.g. Gedzelman, 1988; Heymsfield et al., 1991; Sassen, 1991; Pinto, 1998; Field, 1999; Cober et al., 2001; Larson et al., 2001; Lawson et al., 2001; Sassen et al., 2001; Fleishauer et al., 2002; Hogan et al., 2002; Korolev et al., 2003; Wang et al., 2004; Carey et al., 2008 and others) are observational studies, with far fewer modeling studies (e.g. Starr and Cox, 1985a,b; Liu and Krueger, 1998; Larson et al., 2006, 2007; Larson and Smith, 2009).

Altocumulus clouds tend to form in distinct layers that may be less than 100 m thick. The formation of these clouds is due to a number of different mechanisms including mountain waves, spreading and decaying convective updrafts and anvils, radiational cooling of moist layers, and the lifting of humid air away from its source in sheared environments such as in frontal situations. The existence of liquid water within middle-level clouds results from the balance between the generation of liquid water through vertical motion and its loss due to the Bergeron-Findeisen mechanism. Rimming of liquid drops also represents a loss of liquid water. For conditions of saturation with respect to water, the supersaturation with respect to ice and the Bergeron-Findeisen mechanism both increase with decreasing temperature. The concentration, size, and shape of the ice crystals represent the dominant controls on the Bergeron-Findeisen process.

Heymsfield et al. (1991) sampled two altocumulus clouds and found that the clouds were similar to stratocumulus clouds in that there was extensive cloud top entrainment, a capping inversion and a dry layer above these clouds. Once formed the altocumulus behaved much like stratocumulus with
radiatively-cooled air from near cloud top descending into the cloud layer, although the radiative forcing is enhanced in altocumulus clouds as there is more longwave heating at cloud base. Fleishauer et al. (2002) examined six midlevel clouds that occurred over the Great Plains during the Complex Layered Cloud Experiments (CLEX): a single-layer cloud composed primarily of liquid water, three mixed-phase, single-layer clouds, and two multi-layer mixed-phase clouds. These clouds were found between 2400 and 7000 m, had depths of 500-600 m, and in-cloud temperatures from just below 0 °C to −31 °C. In contrast to Heymsfield et al. (1991), significant temperature inversions or wind shears were not observed in association with any of these clouds, possible reasons for which include: that the clouds did not live long enough to form them, that there were no pre-existing inversions, and as the clouds moved with the wind rather than being anchored to a location, that the wind shear was weak. The absence of inversions and shear-generated turbulence has important implications for dry air entrainment rates.

A common feature of midlevel, single-layer, mixed-phase clouds is the dominant presence of supercooled liquid water in the upper portion of the cloud with ice in the lower regions. Supercooled altocumulus clouds overlaying a deep layer of ice virga are a typical example of such clouds (Hogan et al., 2002; Wang et al., 2004). Temperature inversions at cloud top prevent further vertical development, and a region of ice-saturated relative humidity supports the growth of ice crystals precipitating from the cloud. Two thirds of thirty vertical profiles of mid-latitude, midlevel mixed-phase altocumulus clouds taken over western Nebraska were supercooled, liquid-topped altocumulus clouds with mixed-phase conditions extending to cloud base (Carey et al., 2008). The other ten were glaciated clouds with little evidence of supercooled liquid water. For the mixed-phase clouds, peak liquid water contents occurred in the upper regions of the cloud, typically within tens of meters of cloud top, and ice virga were evident well below cloud base.

### 10.4.2.2. High-Level Clouds

The global coverage of cirrus is of the order of 40% as is evident in Table 10.2 (the CloudSat estimates are lower; see Sassen and Wang (2008) for a discussion on the differences). The greatest coverage occurs in tropical regions, although mid-latitude cirrus are also extensive (Fig. 10.29). Cirrus are by definition semitransparent (Sassen, 2002), and can be difficult to measure observationally. Lidar can detect cirrus with optical depths as low as $10^{-4}$ for wavelengths in the visible part of the spectrum, while satellite sensors need optical depths of ~0.1 in order for their detection. As described by Kärcher and Spichtinger (2009), cirrus clouds are somewhat unique for several reasons. Firstly, cirrus develop in relatively stable thermodynamic environments. Secondly, they are typically associated with relatively high ice nucleation thresholds (supersaturations with respect to ice of tens of percent) and long growth and sublimation time periods due to the low temperatures of the environment in which they form. This results
in regions of long-lived saturations both within and outside of the cloud, as well as difficulties in distinguishing between primary (vapor diffusion processes) and secondary (aggregation and precipitation) ice processes. Thirdly, the transition between clear and cloudy air in cirrus regions is relatively continuous as ice crystals can survive for long periods of time in subsaturated conditions. Ice crystal sedimentation is important in cirrus development and longevity. Finally, rather than originating due to a single mechanism, cirrus clouds can develop from several different mechanisms or sources, and it is these mechanisms that we now examine.

Cirrostratus may form in association with airflow over the warm front in extratropical cyclones. Air flowing over mountains may generate orographic wave clouds of water and/or ice that extend to great heights in the troposphere. Cirrus clouds sometimes form in the vicinity of the jet stream in association with small-scale vertical circulations that develop around the jet. Cirrus also develop in association with closed upper-level lows. Cumulus convection produces significant cirrus clouds through detrainment upon encountering a stable layer, most often the tropopause. TTL cirrus cover large regions near the cold tropical tropopause (Gettleman et al., 2002) and are difficult to detect. Tropical cirrus that occur primarily between 10 and 15 km are optically thicker, contain more condensate, show greater temporal and spatial variability, and tend to be more frequently associated with deep convection at some point in their development than TTL cirrus (Mace et al., 2006b). Finally, condensation trails (contrails) from jet aircraft can produce cirrus clouds.

Cirrus clouds assume a variety of forms, depending on the vertical velocity, wind shear, relative humidity, and static stability. The formation of cirrus clouds is primarily dynamically driven by variations in the vertical wind field on the mesoscale (gravity waves, turbulence) that occur within regions supersaturated with respect to ice. Ordinary cirrus tend to be generated through in situ ice nucleation, whereas ice in anvil and frontal cirrus is typically formed within mixed-phase clouds and then transported aloft. In situ nucleation may then occur following the sedimentation of the previously transported ice. Initial ice crystal concentrations within cirrus are a strong function of vertical velocity and the associated cooling rates. In regions of relatively low cooling rates, heterogeneous ice nucleation may influence cirrus formation. Few numerical modeling studies have been conducted to date that assess the dynamic and aerosol controls of cirrus formation and development.

Anvil cirrus may spread horizontally outward and cover areas many times the size of their generating convective updrafts (Fig. 10.30). The anvils from neighboring cumulonimbus clouds can merge into an extensive stratiform layer in the middle and upper troposphere (Chapter 9). Anvil cirrus occur frequently in the tropics, and are also a significant source of ice in the upper troposphere of the mid-latitudes (Tiedtke, 1993). The mass and number concentration of ice within anvil cirrus are primarily determined by the drop size distributions within the convective updraft. Given the range of convective updrafts that produce
FIGURE 10.30 Satellite photograph of cumulus convection over Florida peninsula and adjacent waters. A thunderstorm in south Florida is producing a massive shield of cirrus clouds.

cirrus anvils, the anvils are characterized by a range of sizes, longevities and optical properties. Thin, TTL cirrus occur most frequently in regions of deep convection (Dessler et al., 2006). These clouds develop in association with the strong updrafts of overshooting tops, which allow for the freezing of liquid particles and the formation of thin cirrus above the level of the main convective detrainment, and hence above typical anvil cirrus.

Contrails are perhaps the only single cloud type that are purely anthropogenically generated (Fig. 10.31). Unlike adiabatic cooling processes that produce middle and high clouds, contrails are generated through the mixing of warm, moist exhaust air with colder, drier environmental air. This mechanism, made possible by the exponential increase of saturation vapor pressure with temperature, is the same mechanism that produces steam fogs when cold, dry air flows over warmer water (Chapter 6). Aspects such as the ambient temperature, pressure and relative humidity with respect to ice (RHI), as well as the specific heat of jet fuel, and the aircraft propulsion efficiency all influence the maximum temperature and minimum relative humidity at which contrails form (Schumann, 1996). In corridors of high aircraft density, individual linear contrails can merge to form a thin blanket of contrail cirrus clouds, a process influenced by the ambient conditions and dynamical processes controlling ice supersaturation. Under suitable conditions contrail cirrus may cover extensive regions (up to 100,000 km²; Duda et al., 2001). It has been suggested that contrails are the reason for the sudden increase in high-level
cloud cover observed over the United States at the start of the jet era (Liou et al., 1990). The radiative forcing of contrails and contrail cirrus is a strong function of their optical depth, which typically varies from 0.1 to 0.5 (Minnis et al., 2005) but can be much smaller or larger.

Dense layers of cirrostratus occur under conditions of gentle, uniform upward motion, saturated air, and high static stability. Under reduced stability and weaker mean upward motion, convection may form cirrus uncinus (mare’s tails), dense patches of cirrus, which produce ice crystals large enough to acquire appreciable terminal velocities. In the presence of wind shear these ice crystal fallstreaks may form trails of considerable length (Fig. 10.32). Because of the high RHI compared to that of water, the crystals may survive falls of more than 5 km through clear air (Hall and Pruppacher, 1976). Sometimes these ice crystals “seed” middle-level supercooled water clouds, thus stimulating the growth of precipitation at the lower levels.

Heymsfield (1975a,b) studied the dynamics and microphysics of cirrus uncinus clouds using aircraft observations. A conceptual model for cases of positive wind shear (west wind increasing with height) throughout the cloud system is presented in Fig. 10.33a. According to this model, clouds are initiated in an updraft that develops in a layer with a nearly dry adiabatic lapse rate. The updraft velocity in this convective head is 1.0-1.5 m s\(^{-1}\), considerably larger than the average updraft velocities in cirrus clouds. As the ice crystals rise, they grow and move downshear relative to their point of origin. When their size and terminal velocity increase and/or they move out of the convective updraft, they fall and form a trail extending upshear of the generating point. When the head forms in a region of negative wind shear, the head curves the opposite way (Fig. 10.33b). The particles grow and are carried downshear from the generating region until they become large enough to fall out of the updraft. In a relative sense, they then move back toward their point of origin until they reach the region of positive shear, when they again move upshear away from

FIGURE 10.31 Photograph of condensation trails. (Photo by R. Anthes.).
the point of origin. The resulting cloud resembles a reversed question mark. An almost infinite variety of trails can be produced depending on the wind shear. Evaporation of the fallstreaks in the lower levels forms part of the transferral of moisture from the upper to lower levels via these cirrus systems.

Atlas et al. (2006) examined the transition of contrails to cirrus uncinus. They noted that downward “pendants” created by the wake dynamics of the aircraft generate the convective turrets that produce the cirrus uncinus. The turrets may grow to 1-2 km in horizontal size. The pendants are characterized by high concentrations of small ice crystals and high IWCs, and grow as a result of heating by longwave radiation from the ground. Mesoscale uncinus complexes have also been observed (Wang and Sassen, 2008).

Lidar is sensitive to optically thin cirrus layers. By comparing visual and lidar detections, two classes of ice clouds were distinguished by Sassen et al. (1989): visible and subvisible clouds. It is not yet clear whether these two cloud classes are due to different formation processes. Thin or subvisible cirrus have been observed in 50% of the lidar observations taken over Taiwan between 1993 and 1995 (Nee et al., 1998), and thin cirrus layers above 15 km occurred in 29% of lidar observations during the Central Equatorial Pacific Experiment (CEPEX) and the Coupled Ocean-Atmosphere Response Experiment of the Tropical Ocean and Global Atmosphere Programme (TOGA COARE). Haladay and Stephens (2009) studied tropical cirrus between 20°N and 20°S using CloudSat radar data and Cloud-Aerosol Lidar and Infrared Pathfinder Satellite...
Observations (CALPISO) lidar data and found that optically thin cirrus covered about 30% of the region, about one third of which occurred as single cloud layers without any clouds below them, and that they demonstrated seasonal variations in association with the annual cycle of convection.

Apart from the identification of visible and subvisible classes, a number of other cirrus classification schemes have also been developed. Keckhut et al. (2006) identified four cirrus classes based on geometric height and thickness, three of which contributed similar proportions (~30%) to the total cirrus detected. The first class consisted of thin clouds occurring slightly above the local tropopause (~11.5 km). These clouds may be related to anvil clouds advected to the mid-latitudes or to moisture transport from the tropical upper troposphere to the stratosphere (Garrett et al., 2004). The second class was also made up of thin cirrus, however they occurred at lower altitudes (~8.6 km). In the third class the clouds were thick (~3 km) and were located between the
other two classes (~9.8 km), just below the local tropopause height. Other cirrus classification schemes have been based on clustering using Meteosat imagery (Desbois et al., 1982) and on shape ratios (Noel et al., 2002).

Two classes of tropical cirrus clouds were identified by Pfister (2001). The first class is linked to the detrainment of particles and water vapor from cumulonimbus clouds, although much of the cirrus at this level may be generated and maintained through processes other than detrainment. The second class occurs in the regions of the TTL, regions characterized by stable conditions. While these cirrus may be associated with deep convection, they do not appear to be directly related to mass detrainment. Pfister (2001) suggested that they are due to isentropic uplift, while Boehm and Verlinde (2000) linked them to vertically propagating Kelvin and gravity waves.

Wang and Sassen (2002) found that most of the cirrus clouds over the ARM SGP site between 1996 and 2000 were typically optically thin (mean of 0.58) with a low IWP (mean of 12.19 g m$^{-2}$), and a general effective radius between 30 and 50 microns. The IWC, general effective radius and extinction coefficients were found to depend strongly on temperature, although these properties did vary significantly at a given temperature thus demonstrating that factors other than temperature are important in controlling cirrus properties. Massie et al. (2002) examined the distribution of tropical cirrus and deep convection and found that approximately one half of their cirrus observations occurred in association with deep convection, while the other half developed due to in situ processes.

Mace et al. (2006b) used a year of observations to compare the properties of tropical cirrus (excluding the TTL type) that occurred over the ARM sites, Manus and Nauru islands, in the tropical West Pacific. Manus is located within the center of the Pacific warm pool, a region of frequent and widespread deep convection, whereas Nauru is located on the eastern edge of the warm pool where deep convection is less frequent. Cirrus above 7km over Manus occurred 48% of the time compared with only 23% of the time over Nauru, and were thicker and warmer on average. Less than half of the cirrus observed at both islands could be traced to deep convection within the past 12 hours indicating that cirrus in the Tropics must evolve into structures that are maintained by other dynamical processes that support their longevity. A clear temporal evolution in cirrus properties was evident, with both the radar Doppler moments and the IWP decreasing with increasing cloud age. The cirrus properties also appeared to be sensitive to the properties of the deep convective source. Over Manus during the boreal winter, the cirrus had higher water contents and higher small particle concentrations compared to the summer cirrus, being due to the fact that the trajectories associated with cirrus during the winter originate to the south and east of the island where deep convection occurs frequently in association with the winter monsoon, whereas during the boreal summer the trajectories originate in the maritime regions to the north and east of the island where convection is less frequent.
Properties of cirrus are closely tied to the distance of the cirrus from the parent convection, with larger crystals being found nearer cloud base and closer to the connective core (McFarquhar and Heymsfield, 1996). Hoyle et al. (2005) noticed regions of high number concentrations of ice particles (as high as 50 cc\(^{-1}\)) embedded within more widespread regions of lower concentrations (0.1-50 cc\(^{-1}\)) from measurements made of in situ-formed cirrus (without convective or orographic influences). They attributed these observations to small-scale temperature fluctuations associated with gravity waves, mechanical turbulence or other small-scale air motions. The waves were found to have frequencies of 10 hr\(^{-1}\) and peak-to-peak amplitudes of 1-2 K, with instantaneous cooling rates of up to 60 K hr\(^{-1}\). Hoyle et al. suggested that the properties of young, in situ forming cirrus clouds are primarily determined by homogeneous freezing forced by a range of small-scale fluctuations (on the period of several minutes) with moderate to high cooling rates (1-100 K hr\(^{-1}\)).

Garrett et al. (2005) examined a case study of deep convection and its associated anvil cirrus over Florida during the Cirrus Regional Study of Tropical Anvils and Cirrus Layers—Florida Area Cirrus Experiment (CRYSTAL-FACE) (Jensen et al., 2004). They found that the detrained cloud mass was separated into two vertical layers, a cirrus anvil with cloud top temperatures of \(\sim -45 \, ^{\circ}C\) and a thin tropopause cirrus (TTC) layer with temperatures of \(\sim -70 \, ^{\circ}C\) which had similar dimensions to the anvil layer, and which lay \(\sim 1.5 \, \text{km}\) above the anvil layer. The anvil layer was optically thick, while the second layer only had a visible optical depth of 0.3 and IWCs of \(\sim 1 \, \text{mg m}^{-3}\) in spite of being 1.5 km thick. The unimodal size distribution of the TTC was characterized by small ice particles (effective radius of 4-10 microns), and the microphysical properties remained relatively uniform over time. This may have been due to the fact that the lower anvil shielded the TTC from surface infrared radiation thereby cooling it (Hartmann et al., 2001). Garrett et al. suggested that these TTCs might evolve into widespread subvisible cirrus over low-latitude regions.

In the anvil, the ice crystals larger than 50 microns were found to aggregate and precipitate out, thus resulting in reduced IWCs and an ice crystal population dominated by smaller sizes. The anvil cloud thinned from \(\sim 2.5 \, \text{to} \, 0.5 \, \text{km}\) during this time. The top-of-atmosphere (TOA) radiative forcing ranged from strong cooling near the leading edge of the anvil and tended toward zero as the anvil thinned downstream of the convective core, being influenced both by a decrease in IWP and an increase in effective radius. The TTC on the other hand made little contribution to the TOA except when the anvil became transparent. The anvil cirrus appeared to spread laterally due to the radiatively-driven pressure gradients at the cloud boundaries rather than convection through the entire layer. Finally, while the TTC layer also spread out, it did not dissipate, possibly due to the shielding provided by the anvil cirrus.

Cirrus cloud structures play important roles in satellite retrievals, radiative calculations and cloud dynamic forcing (e.g. Quante and Starr, 2002; Hogan and Kew, 2005). These structures are difficult to analyze as they are typically
non-stationary and comprise a number of different length scales. For example, cirrus uncinus cells (∼1 km) may be made up of groups of smaller updrafts (∼100 m) and can also occur in larger uncinus complexes (∼10-100 km) (Sassen et al., 1989). Various dynamic phenomena may also be observed in cirrus clouds, such as small convective cells, gravity waves (wavelengths of 2-9 km), quasi-two-dimensional waves (wavelengths of 10-20 km) and larger two-dimensional mesoscale waves (wavelengths of ∼100 km) (Gultepe and Starr, 1995). While most cirrus layers are horizontally inhomogeneous, locally periodic structures are common in cirrus and include cirrus uncinus cells at cloud top, cirrus mammata at cloud base, and KH waves embedded within the cirrus (Sassen et al., 2007). Even though these structures have different formation mechanisms, they have similar average length scales (1-10 km), can coexist, and tend to occur with greatest frequency in the lower regions of the cloud layer.

10.4.3. Environmental Properties of Middle- and High-Level Clouds

The upper troposphere is often supersaturated with respect to ice, the evidence of which is the large spatial scale of anvil cirrus and the development of contrails in the absence of other upper level clouds. Starr and Cox (1980) examined temperature, moisture, and wind soundings associated with more than 3600 cloud cases to determine the characteristics of the environment of middle and high clouds. Knowledge of these characteristics is useful to develop methods of parameterizing clouds in large-scale models as a function of the large-scale thermodynamic and dynamic variables predicted by the model. The parameters examined by Starr and Cox included static stability, defined by

\[ \sigma \equiv \partial \theta / \partial z, \]  

(10.11)

vertical wind shear,

\[ S \equiv |\partial V / \partial z| \]  

(10.12)

and the Richardson number,

\[ R_i = (g/\theta)(\sigma / S^2), \]  

(10.13)

in addition to the mean temperature and relative humidity in the cloud layer.

A common property of most cloud observations was that the lapse rate was rarely saturated adiabatic. Other properties of the environment showed considerable variation from case to case, indicating the difficulty of parameterizing the cloud effects in large-scale models. There were statistical differences between the characteristics of thick cloud layers (defined as extending through a depth of greater than 50 mb) and thin clouds (50 mb
TABLE 10.3 Characteristics of thick and thin middle- and high-level clouds as deduced from radiosonde ascents

<table>
<thead>
<tr>
<th>A. Thick clouds (depth greater than 50 mb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Static stability:</td>
</tr>
<tr>
<td>Decreases with height from subcloud layer to above cloud layer, with a typical range of 5 K km$^{-1}$ in the subcloud layer to 3 km$^{-1}$ near the cloud top.</td>
</tr>
<tr>
<td>Vertical wind shear:</td>
</tr>
<tr>
<td>Usually positive (increasing wind speed with height): magnitude ranging from about 5 m s$^{-1}$ km$^{-1}$ in summer to 6.5 m s$^{-1}$ km$^{-1}$ in winter.</td>
</tr>
<tr>
<td>Richardson number:</td>
</tr>
<tr>
<td>Average values of $R_i$ throughout cloud layer 12-18; very small percentage (about 15%) of soundings showed $R_i$ less than 1.0.</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>B. Thin clouds (depth less than or equal to 50 mb)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Static stability:</td>
</tr>
<tr>
<td>Quite variable, but on average greatest stability (5.5 K km$^{-1}$) above cloud layer and least stability (3.5 K km$^{-1}$) in and below cloud layer; stable layer on top may correspond to tropopause.</td>
</tr>
<tr>
<td>Wind shear:</td>
</tr>
<tr>
<td>No significant difference from thick clouds.</td>
</tr>
<tr>
<td>Richardson number:</td>
</tr>
<tr>
<td>Average values slightly greater (18-24) than for thick clouds.</td>
</tr>
</tbody>
</table>

Source: Adapted from Starr and Cox (1980).

thickness or less). In general, thick clouds were associated with frontal circulations and cyclonic storms and thin clouds were not. The general characteristics of thick and thin clouds are summarized in Table 10.3.

Two conclusions may be drawn from the summary of average cloud conditions in Table 10.3. First, the layers are generally statically stable in an absolute sense, with lapse rates less than the saturation-adiabatic rates with respect to either water or ice. The fact that thick clouds are generally above stable layers indicates that they are associated with upper-level fronts. By contrast, thin clouds are usually below stable layers, which indicates that they are located below the tropopause. A second conclusion from Table 10.3 is that shear-induced turbulence (Kelvin-Helmholtz instability) is not likely in the vicinity of most middle- and high-level cloud systems. A necessary condition for Kelvin-Helmholtz instability is that the Richardson number be less than 1/4. As shown in Table 10.3, the average values of $R_i$ are much greater than this value. The coarse vertical resolution used to evaluate $R_i$ in this study may bias the estimates of $R_i$ toward higher values; it is possible that considerably smaller values of $R_i$ exist locally in thin layers where strong destabilization associated with radiative effects is important. An exception to this conclusion occurs in shallow layers in the vicinity of jet streams, where mesoscale vortices produce turbulence and extensive bands of cirrus.

Cirrus clouds develop primarily in ice-supersaturated regions which, in the upper troposphere at mid-latitudes, are typically $\sim$150 km in extent, although
they may be larger (∼1000 km), and are ∼0.5-1.5 km deep, although they may be as deep as 3-5 km. Three mechanisms appear to play a role in enhancing the RHI including (1) adiabatic vertical motion on a range of spatial and temporal scales; (2) turbulent mixing of air parcels due to wind shear or dynamic instabilities; and (3) diabatic effects associated with shortwave and longwave radiation. Turbulent mixing and diabatic effects typically occur on timescales too long to initiate cloud formation, but appear to influence the cloud life cycle once formed (Kärcher and Spichtinger, 2009). While variations in water vapor affect the formation of cirrus, this is of secondary importance to changes in temperature and the associated changes in RHI. Cirrus clouds thus differ from lower tropospheric clouds where water vapor variations are more important. The properties of cirrus appear to be only weakly dependent on rising motion on the synoptic scale, thus suggesting that vertical velocity on the small scale may play a more significant role in determining their properties (Mace et al., 2001; Quante and Starr, 2002). Kärcher and Ström (2003) demonstrated that the properties of cirrus cannot be simulated when only making use of synoptic cooling rates, and that mesoscale variability in vertical velocity and the associated temperature variations are the primary controllers of the microphysical characteristics of cirrus. Such mesoscale variations in vertical velocity and temperature are associated with gravity waves produced by factors such as orographic forcing, convection, baroclinic instability and geostrophic adjustment.

As cirrus characteristics and properties vary on a seasonal basis, this demonstrates their dependency on synoptic patterns (Mace et al., 2006a). Sassen and Campbell (2001) identified three major synoptic patterns of importance: a zonal jet stream flow, split-jet flow and strong amplitude ridges. Cirrus in the tropics appear to be associated with deep convection at ∼50% of the time, but the origin of the other 50% is not known. In the mid-latitudes, cirrus characteristics appear to be a stronger function of the large-scale vertical motion within cloud systems (Mace et al., 2006a). Deng and Mace (2008) compared the properties of mid-latitude and tropical cirrus clouds and found that the cirrus clouds in the tropics had larger particle sizes, greater ice masses, were more likely to be associated with ascending air motions, and were colder and higher in altitude. The mid-latitude cirrus showed strong seasonal variations whereas interannual variations (such as variations due to ENSO) were stronger than seasonal variations for tropical cirrus. The seasonal variations in the tropics and subtropics are associated with convective cycles and the movement of the ITCZ (Stubenrauch et al., 2006). Cirrus clouds were found to be more frequent and thicker during the summer than in winter in the mid-latitudes. Diurnal variations in cirrus were observed both in the tropics and the mid-latitudes, although tropical cirrus showed a greater variation. Tropical cirrus occurred higher in the atmosphere and were thicker in the morning and then decreased in the afternoon. In the mid-latitudes there was no obvious diurnal variation in winter, however, during the summer, cirrus were found at higher altitudes and were thicker during the afternoon, which is indicative of the development of deep convection and
the associated spreading of anvil. In the western tropical Atlantic ITCZ, dual maxima (morning and late afternoon) have been found, the causes of which are not well understood.

Mace et al. (2006a) observed a seasonal variation in the macroscale properties of cirrus over the ARM SGP site. A minima in frequency occurred in late summer, while the maxima existed in late winter and early spring when synoptic-scale systems and strong jet streams are associated with deep convection. Cirrus occurred over lower-level clouds 1/3 of the time, and were more frequent in winter. Cirrus tended to occur within regions of upper-troposphere humidity maxima and just downstream of a peak in upper-troposphere vertical motion. Those cirrus that formed within large-scale ascent regions upstream of mid-tropospheric ridges had higher water content than those forming in regions of large-scale subsidence downwind of the ridge axis. In summer, while cirrus still appeared to be associated with large-scale ascent the anomalies were less distinct, suggesting that warm season cirrus tend to be more frequently associated with detrainment from deep convection.

The condensate detrainment associated with deep convection first forms cirrostratus and then thin cirrus. It has been estimated that it takes ~6-12 hours for the development of cirrostratus, followed by ~1 day for the transition to thin cirrus. The cirrus amount has an e-folding time of ~5 days, and the mean lifetime of convectively driven cirrus is ~10 days (Luo and Rossow, 2004). Contrails can form and persist in environments that are saturated with respect to ice, whereas natural cirrus require high ice supersaturations to form. Thus contrails can occur in environments that may otherwise be cloud free. The life cycle of contrail cirrus is poorly understood.

10.4.4. Microphysical Properties of Middle- and High-Level Clouds

10.4.4.1. Middle-Level Clouds

Measurements of the microphysical properties of middle clouds are relatively sparse. In situ measurements of droplets made by Borokvikov et al. (1963) showed that droplets averaged about 10 microns in size. A more detailed investigation into the microphysical characteristics of a 300 m deep altocumulus sampled between −9° and −12 °C showed that the mean droplet size ranged from 7 to 10 microns, droplet concentrations averaged ~ 300 cm$^{-3}$, and the LWC ranged from 0.03 to 0.09 g m$^{-3}$ (Herman and Curry, 1984). Retrieved LWPs and effective radii for the supercooled liquid regions of altocumulus of 15 g m$^{-2}$ and 6 microns, respectively, have also been observed (Wang et al., 2004).

Using observations and simple model calculations of droplet growth in altocumulus clouds Heymsfield et al. (1991) observed typical droplet growth patterns beginning with a rapid increase in droplet concentrations and mean diameters immediately above cloud base, followed by nearly constant peak
drop concentrations and diameters that increase in size much more slowly, and finally, a rapid decrease in mean diameters and cloud droplet concentrations near cloud top due to mixing with the dry, overlying air. Peak supersaturations are much higher than typically calculated for cumulus clouds due to low droplet concentrations and comparatively low droplet growth rates relative to the vapor supply rate. Few, if any, ice crystals were observed in the altocumulus clouds they examined, which they attribute to the absence of ice nuclei.

Observations of the evolution of the size spectra of ice crystals larger than 800 microns within an altostratus cloud associated with a cold front showed variations to occur on horizontal scales of \( \sim 5 \) km (Field, 1999). Between \(-40 \) °C and \(-20 \) °C, particle growth was dominated by diffusional growth, although aggregation was also evident. Between \(-20 \) °C and \(-10 \) °C growth was dominated by aggregation. Rauber and Grant (1986) observed that supercooled water can be produced in regions in the cloud where the condensational rate exceeds the diffusional rate of ice crystals.

In the CLEX study of single and multiple layer mixed-phase midlevel clouds, Fleishauer et al. (2002) noted in the thin, mixed-phase single-layer clouds that the LWC increase with altitude, whereas the IWC maximizes in the mid- to lower regions of the cloud. This is in contrast to the multilayered systems in which ice was more evenly distributed throughout the cloud layer. In the thin, single-layer clouds, drop sizes were found to increase with increasing altitude. Peak LWCs were \( 0.35 \) g m\(^{-3}\) with mean values ranging from \( 0.01 \) to \( 0.15 \) g m\(^{-3}\). The seeder-feeder mechanism was examined as a potential mechanism causing the differences in vertical ice distribution between the single layer and multilayer cases. While some of the ice crystals had habits originating from colder regions than where they were observed, and ice crystals were detected in the clear air between the cloud layers, the number of crystals between layers was small, thus suggesting that this mechanism was not significant. Finally, their observations showed a poor correlation between IWC and temperature.

Larson et al. (2001) examined the causes of the dissipation of one of the altocumulus clouds observed by Fleishauer et al. (2002) during CLEX. An analysis of the liquid water budget of this cloud showed that the cloud did not dissipate due to precipitation fallout. Rather, the largest contribution to the liquid water decay was subsidence drying. The net effect of radiative transfer on this cloud was unclear. In a subsequent modeling study of the same case, Larson et al. (2006) made use of three-dimensional, LES simulations in order to assess the role played by radiative heating or cooling, large-scale subsidence, diffusional growth of ice and turbulent mixing of dry air into the cloud in the dissipation time of the midlevel clouds. They found that subsidence and ice diffusional growth had non-positive contributions at all altitudes, thus leading to a strong decrease in liquid water. They also estimated from their sensitivity tests that increasing the solar zenith angle by 1 (the maximum possible change) reduces the cloud lifetime by the same amount as increasing the subsidence
velocity by 1.21 cm s\(^{-1}\), increasing the ice number concentration by 781 m\(^{-3}\) or decreasing the total water mixing ratio above cloud by 0.597 g kg\(^{-1}\) for this particular case.

**Hobbs and Ragno (1985)** observed two types of altocumulus (castellanus and floccus) and found that the LWC maxima occurred at cloud top, together with a small number of ice particles, while large concentrations of ice particles occurred in the lower regions of the cloud, together with little or no LWC. Carey et al. (2008) also found peak LWCs at or near cloud top and peak IWCs in the lower half of the cloud in their altocumulus study. They also observed that as the long-lasting (9-10 hours) cloud field dissipated, the altocumulus cloud appeared to “deglaciate” in that while the LWC remained the same, the IWC decreased dramatically as the cloud progressed from the early through to the dissipation stage. A similar “deglaciation” was also noted in another mixed-phase study by Kankiewicz et al. (2000). This deglaciation process was observed to occur even while there was still significant supercooled water present, and may have resulted due to decreases in the ice nuclei supply. Ice nuclei measurements were not however made during the field campaign. Larson and Smith (2009) conducted LES simulations of three thin, midlevel layer clouds in order to investigate the sensitivity of the glaciation rate of altocumulus clouds to habit type. They found that the relationships were complex, but that, for the cases they examined, dendrites tend to glaciate altocumulus clouds more rapidly than plates.

### 10.4.4.2. High-Level Clouds

The size, shape and concentration of ice crystals in cirrus clouds strongly influence the earth’s radiation budget. Using a climate model, Kristjánsson et al. (2000) showed that variations in ice crystal habit and size may have a significant influence on the simulated climate change. Stephens et al. (1990) concluded that our understanding of the impacts of cirrus on climate is limited by our knowledge of the relationship between ice crystal size and shape, and the broad radiative properties of cirrus. Measurements of ice crystal sizes and shapes are also important to obtain accurate remote sensing products. One of the primary reasons for our lack of understanding of the processes involved in ice crystal formation within cirrus is the difficulty in observing the formation of cirrus in situ. Cirrus clouds are cold clouds that form high in the troposphere, especially in the Tropics. Also, unlike the formation of ice particles in cumulus clouds and wave clouds, cirrus clouds do not have an obvious initial stage, and thus it is difficult to place aircraft in the right locations at the right time to observe ice crystal formation.

Aircraft observations of middle- and high-level clouds have revealed much about their structure and microphysical characteristics. As summarized in Table 10.4, cirrus clouds contain ice crystals typically 0.5 mm in length and concentrations that vary widely from 10\(^4\) to 10\(^6\) m\(^{-3}\). The terminal velocity of the ice crystals is about 0.5 m s\(^{-1}\), which is generally of the same order as
TABLE 10.4  Microphysical characteristics of Middle and High Cirrus Clouds

<table>
<thead>
<tr>
<th>Property or variable</th>
<th>Value</th>
<th>Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Concentration of ice crystals (m⁻³)</td>
<td>10⁵-10⁶</td>
<td>Braham and Spyers-Duran (1967)</td>
</tr>
<tr>
<td></td>
<td>2 × 10⁵-5 × 10⁵</td>
<td>Heymsfield (1975a)</td>
</tr>
<tr>
<td></td>
<td>1.0 × 10⁴-2.5 × 10⁴</td>
<td>Heymsfield and Knollenberg (1972)</td>
</tr>
<tr>
<td></td>
<td>6.0 × 10⁵-38.0 × 10⁵</td>
<td>Ryan et al. (1972)</td>
</tr>
<tr>
<td></td>
<td>2.0 × 10⁴-8.0 × 10⁴</td>
<td>Houze et al. (1981)</td>
</tr>
<tr>
<td></td>
<td>1.0 × 10⁴-5.5 × 10⁵</td>
<td>Churchill and Houze (1984)</td>
</tr>
<tr>
<td>Length of crystals</td>
<td>Up to 0.17 mm</td>
<td>Braham and Spyers-Duran (1967)</td>
</tr>
<tr>
<td></td>
<td>0.6-1.0 mm</td>
<td>Heymsfield and Knollenberg (1972)</td>
</tr>
<tr>
<td></td>
<td>0.35-0.9 mm</td>
<td>Heymsfield (1975a)</td>
</tr>
<tr>
<td>Terminal velocity of crystals</td>
<td>Typically 50 cm s⁻¹; max 120 cm s⁻¹;</td>
<td>Heymsfield (1975a)</td>
</tr>
<tr>
<td>Ice-water content</td>
<td>0.15-0.25 g m⁻³</td>
<td>Heymsfield and Knollenberg (1972)</td>
</tr>
<tr>
<td></td>
<td>0.15-0.30 g m⁻³</td>
<td>Heymsfield (1975a)</td>
</tr>
<tr>
<td></td>
<td>0.10-0.50 g m⁻³</td>
<td>Rosinski et al. (1970)</td>
</tr>
<tr>
<td>Precipitation rate</td>
<td>0.5-0.7 mm h⁻¹</td>
<td>Heymsfield and Knollenberg (1972)</td>
</tr>
<tr>
<td>Updraft velocity</td>
<td>1.0-1.5 m s⁻¹ (in cirrus uncinus)</td>
<td>Heymsfield (1975a)</td>
</tr>
<tr>
<td></td>
<td>2.0-10.0 cm s⁻¹ (warm front overrunning)</td>
<td>Heymsfield (1977)</td>
</tr>
<tr>
<td></td>
<td>25-50 cm s⁻¹ (closed low aloft)</td>
<td>Heymsfield (1975a)</td>
</tr>
<tr>
<td>Lifetime individual cloud</td>
<td>15-25 min</td>
<td>Heymsfield (1975a)</td>
</tr>
</tbody>
</table>

the updraft speed. The water content of cirrus clouds is typically 0.2 g m⁻³. Observations have shown that IWC is strongly dependent on the vertical velocity and the temperature. Figure 10.34 shows plots of observed IWCs versus temperature for vertical velocities ranging from 1 to 50 cm s⁻¹. The observations were made under various synoptic situations, including warm-frontal overrunning, warm-frontal occlusions, closed lows aloft, and jet stream cloudiness (Heymsfield, 1977). The data shown in Fig. 10.34 indicate, for a
given updraft speed (W), a nearly exponential increase of water content with increasing temperature, from low values of less than $10^{-3}$ g m$^{-3}$ at temperatures below $-50$ °C to maximum values around 0.3 g m$^{-3}$ at temperatures in the range of 0 to $-10$ °C. At a given temperature (e.g. $-30$ °C), the water content increases from $10^{-2}$ g m$^{-3}$ for an updraft of 0.01 m s$^{-1}$ to $2.0 \times 10^{-1}$ g m$^{-3}$ for an updraft speed of 0.5 m s$^{-1}$. The empirical equation fitting the data in Fig. 10.34 is

$$IWC = 0.072 W^{0.78} \exp[-0.01 W^{0.186} (-T)^{1.59} W^{-0.04}]. \quad (10.14)$$
The total ice-crystal concentration also shows a strong dependence on temperature and vertical velocity. As shown in Fig. 10.35, the concentration varies from less than $10^{-1}$ liter$^{-1}$ (100 m$^{-3}$) for weak (0.01 m s$^{-1}$) updrafts and temperatures below $-50$ °C to 100 liter$^{-1}$ ($10^5$ m$^{-3}$) for temperatures around $-10$ °C and updraft speeds of 0.5 m s$^{-1}$. Observations also indicate a relationship between crystal size and IWC. Figure 10.36 is a plot of observed mean and maximum crystal lengths as a function of IWC. The maximum lengths in particular, show an increase with increasing IWC, from about 0.5 mm at $10^{-4}$ g m$^{-3}$ to 5 mm at 1 g m$^{-3}$.

A property of precipitating cloud systems having practical consequences is the close relationship between the precipitation rate $R$ and IWC. Figure 10.37 shows this strong positive correlation, which is well represented by

$$R = 3.6(IWC)^{1.17},$$

(10.15)

where IWC is expressed in g m$^{-3}$ and $R$ is in mm h$^{-1}$. A very similar expression
was derived independently by Sekhon and Srivastava (1970),

$$R = 5.0(IWC)^{1.16}. \quad (10.16)$$

This empirical relationship is useful because radar reflectivity $Z$ is a measure of IWC (Fig. 10.38),

$$Z = 750(IWC)^{1.98}, \quad (10.17)$$

where $Z$ is expressed in units of mm$^6$ m$^{-3}$. Equations (10.16) and (10.17), or other similar empirical expressions, can be used with radar measurements to estimate precipitation rates from cirrus clouds.
Aircraft observations of low-latitude anvil cirrus (e.g. Griffith et al., 1980; Knollenberg et al., 1993; McFarquhar and Heymsfield, 1997; Heymsfield et al., 1998) show that the IWC varies horizontally, tends to decrease with height and can range from \( \sim 0.01 \, \text{g m}^{-3} \) to more than \( \sim 0.1 \, \text{g m}^{-3} \). Observations from field campaigns demonstrate that the size and shape of ice particles in cirrus clouds vary significantly, both spatially and temporally. Processes such as aggregation and sedimentation result in large ice crystals (1000 microns or larger) near the cloud base. Aggregation does not always occur however, and appears to be less frequent than in other clouds. It is sometimes difficult to distinguish aggregation from the complex crystals arising out of diffusional growth processes. Near cloud top, concentrations of ice crystals smaller than 100 microns may be greater than 100 cc\(^{-1}\). Measurements in cirrus that formed independent of convective or orographic influences show number concentrations as high as 50 cc\(^{-1}\) within regions of concentrations of 0.1-50 cc\(^{-1}\). Others have shown number concentrations of the order of 0.1 cc\(^{-1}\) with values as high as 10 cc\(^{-1}\), especially in younger cirrus (Mace et al., 2001; Kärcher and Ström, 2003). Optically thin cirrus may have concentrations of the order of 0.1 cc\(^{-1}\) (Mace et al., 2001).
Microphysical properties of cirrus clouds appear to be weakly correlated with ambient environmental conditions (Korolev et al., 2001), although particles appear to become more regularly shaped as the temperature and supply of water vapor decreases (Magano and Lee, 1966). The size and structure of cirrus ice crystals determines their sedimentation velocity, which in turn influences cloud lifetime and the humidity of the upper troposphere. If the total available water is spread over a smaller number of large ice crystals then the cloud dehydrates more rapidly due to greater sedimentation rates than if the same total water was spread over a larger number of smaller ice crystals. The microphysical properties of contrail cirrus are likely to differ from those of natural cirrus, at least during the initial stages of development, due to the nucleating properties of aircraft particles.

The crystal habit and size has important radiative implications. Heymsfield and Platt (1984) observed that the predominant crystal types for temperatures between \(-20\,^\circ\text{C}\) and \(-40\,^\circ\text{C}\) were polycrystalline forms, with some columns,
plates and bullets. Between $-40 \, ^\circ\text{C}$ and $-50 \, ^\circ\text{C}$, hollow columns were predominantly observed, and at temperatures below $-50 \, ^\circ\text{C}$, hollow and solid columns, together with some hexagonal plates and thick plates were primarily seen. Bullet rosettes were also sometimes found, although they tended to be infrequent. More recent observations of ice crystals within cirrus have shown that mid-latitude and high-latitude cirrus may contain higher concentrations of rosette-shaped crystals than previously recorded.

Lawson et al. (2006) analyzed data from 22 mid-latitude cirrus clouds. The cloud temperatures of these cirrus clouds ranged from $-28 \, ^\circ\text{C}$ to $-61 \, ^\circ\text{C}$. They showed that the ice particle size distributions of cirrus are primarily bimodal, with a maximum in mass, area and number concentration near 30 microns, and the second mode, a smaller maximum, occurring between 200-300 microns. Rosette shapes were found to be the predominant particle shape. They also observed that crystal habit categories are a function of particle size. Spheroids dominate the particle size distribution for sizes less than 30 microns; between 30 and 150 microns, small irregular crystals contribute to most of the mass with smaller contributions from columns in this size range; budding rosettes contribute $\sim 25\%-40\%$ of the mass for sizes between 75 and 300 microns, and rosette shapes contribute more than 80% of the mass for crystal sizes larger than 200 microns. The percentage contribution of rosette shapes increases with increasing temperature, from $\sim 40\%$ at $-50 \, ^\circ\text{C}$, to 55% at $-40 \, ^\circ\text{C}$ to 80% at $-30 \, ^\circ\text{C}$. The trend in surface areas is similar. Rosette shapes made up 50% of the surface area and mass of ice particles greater than 50 microns in size, with irregular shapes making up 40% of the remaining 50%, columns and spheroidal shapes constituting several percent, and plates accounting for 1% of the total mass. Particles that are less than 50 microns in size contribute $\sim 99\%$ of the total number concentration, $\sim 40\%$ of the mass and $\sim 70\%$ of the shortwave extinction in these mid-latitude cirrus. Also, the average particle number concentrations was found to be of the order of 1 cc$^{-1}$, although they were greater than 5 cc$^{-1}$ occasionally. Table 10.5 summarizes various ice crystal characteristics for three different temperature zones. Finally, the microphysical properties of these cirrus clouds were found to be similar to wave clouds thus suggesting that ice particles in cirrus first convert to liquid water and/or solution drops before freezing.

The results of Lawson et al. (2006) are similar to those of Heymsfield and Miloshevich (1995) regarding bimodal size distributions and particle shapes, however, the number concentrations observed by the latter are one to two orders of magnitude lower. Also, Heymsfield and Miloshevich found that deep cirrus clouds have a distinct structure with an ice crystal generation region near cloud top that contains the highest concentrations of small particles, an ice-saturated region below this that supports the growth of ice particles, and finally a lower region in which ice particles sublimate. However, Lawson et al. suggest that cirrus clouds are much more complex in that there is a lack of consistency in particle habits as a function of position in cloud, particles that are nucleating,
| TABLE 10.5 | Means, standard deviations and extreme values of microphysical properties from flights through 22 cirrus clouds. The measurements are separated into three temperature zones and $N$ is the number of legs flown in each temperature zone |
|-----------------|-----------------------------------------------------------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|-----------------|
|                | Mean              | Std dev           | Max              | Min              |
| $\text{IWC (g m}^{-3}\text{)}$ | 0.005             | 0.007             | 0.024            | 0.001            |
| Concentration ($\text{L}^{-1}$)    | 846               | 853               | 3679             | 36               |
| Extinction ($\text{km}^{-1}$)     | 0.464             | 0.585             | 3.00             | 0.023            |
| $R_{\text{eff}}$ ($\mu\text{m}$) | 15.4              | 4.7               | 26.0             | 8.1              |
| Radar reflectivity (dBZ)          | $-27.5$           | $-21.6$           | $-13.8$          | $-51.6$          |
| $\text{IWC (g m}^{-3}\text{)}$ | 0.013             | 0.017             | 0.077            | 0.001            |
| Concentration ($\text{L}^{-1}$)    | 968               | 951               | 3817             | 30               |
| Extinction ($\text{km}^{-1}$)     | 0.920             | 1.060             | 4.779            | 0.017            |
| $R_{\text{eff}}$ ($\mu\text{m}$) | 21.3              | 7.10              | 41.46            | 9.99             |
| Radar reflectivity (dBZ)          | $-21.3$           | $-18.59$          | $-11$            | $-53.5$          |
| $\text{IWC (g m}^{-3}\text{)}$ | 0.037             | 0.038             | 0.125            | 0.002            |
| Concentration ($\text{L}^{-1}$)    | 2170              | 2159              | 8267             | 81               |
| Extinction ($\text{km}^{-1}$)     | 2.404             | 2.663             | 11.094           | 0.140            |
| $R_{\text{eff}}$ ($\mu\text{m}$) | 26.5              | 7.5               | 43.1             | 13.0             |
| Radar reflectivity (dBZ)          | $-15.5$           | $-14.8$           | $-8.8$           | $-39.1$          |

Source: After Lawson et al. (2006)

Growing and sublimating can be found throughout the cloud, and almost any of the crystal shapes can be found either at cloud top or cloud base.

Wu et al. (2000) conducted numerical simulations in order to investigate radiative effects on the diffusional growth of ice particles in cirrus clouds. They found such effects to be both significant and complex. Even in a radiatively-cooled environment, ice particles may undergo radiative warming as the net radiation received by an ice particle depends on the emission from the particle, and the local upwelling and downwelling radiative fluxes. Radiative warming of an ice particle restricts the diffusional growth of the particle. For sufficiently large ice crystals, the associated high radiative warming produces surface ice saturation vapor pressures that are so large as to produce sublimation of the crystals, while smaller particles may grow by vapor deposition under such conditions. Radiative cooling of ice particles on the other hand enhances ice production. For the cirrus case study that Wu et al. (2000) simulated they found that the overall influence of radiative feedbacks was to significantly reduce the total ice mass of the cloud, especially the production of large ice crystals. In
subsequent LES simulations, Cheng et al. (2001) simulated both a thin and a thick cirrus event. They found that, while latent heat release within the thin cirrus clouds was insufficient to generate positive buoyancy, positively buoyant cells could be produced within thick cirrus layers. The pressure perturbations induced by the associated updrafts subsequently affected the cloud evolution. Cheng et al. (2001) also found that radiative cooling and latent heating were of similar magnitudes in well-developed, deep cirrus layers and that latent heating thus needs to be taken into account when developing cirrus models. Also, while probability density functions (PDFs) of the vertical velocity in the more radiatively-driven, thin cirrus were approximately Gaussian in distribution, those for the deep, more convectively unstable cirrus were multimodal and broad. These variations in cirrus characteristics make the parameterization of such clouds within numerical models a very difficult task.

10.4.5. Radiative Properties of Middle- and High-Levels Clouds

Clouds affect both solar (shortwave) and infrared (longwave) radiation. Shortwave cloud radiative forcing depends on the LWP and IWP, as well as on cloud particle size and habit, and is usually negative at the TOA as clouds reflect sunlight. Longwave cloud radiative forcing at the TOA depends on the cloud top temperature. This forcing is usually positive, and largest for clouds with high cold cloud tops. The net cloud radiative forcing at the TOA is then dependent on the difference between the longwave and shortwave radiative forcing. Therefore high clouds in the Tropics can have either strong warming or cooling effects on the TOA radiation budget depending on their optical depth, whereas low-topped clouds always have a net radiative cooling effect.

Many previous studies of cloud radiative forcing have focused on the effects of clouds at the TOA however, as pointed out by Stephens (2005), important feedbacks may also exist in association with the separate effects of clouds on the atmospheric and surface radiation budgets. Relative to clear skies, clouds at low latitudes warm the atmosphere due to increased IR absorption and emission at colder temperatures, and cool the surface due to the reflection of solar radiation back to space. These two effects are often largely reciprocal as observed at the TOA. Clouds at high latitudes tend to affect the radiation balance in a reverse manner to those at low latitudes. The effect of clouds on the atmospheric radiation balance is dependent on the vertical location of the clouds. High cold clouds tend to warm the atmospheric column (relative to clear skies) especially at low latitudes, whereas low clouds enhance the cooling of the atmosphere, especially at high latitudes (Slingo and Slingo, 1988).

Several energy balance studies have been conducted linking ISCCP cloud properties to radiative fluxes at the TOA, the surface and within the atmosphere (e.g. Okhert-Bell and Hartmann, 1992; Chen et al., 2000). Okhert-Bell and Hartmann correlated ISCCP and Earth Radiation Budget Experiment (ERBE) data in order to assess what types of clouds contribute most significantly
TABLE 10.6 The contributions to the longwave, shortwave and net cloud forcings by the five different cloud classes identified by Okhert-Bell and Hartmann (1992) The cloud amount (N) for each cloud type is also given

<table>
<thead>
<tr>
<th>Type</th>
<th>Type 1</th>
<th>Type 2</th>
<th>Type 3</th>
<th>Type 4</th>
<th>Type 5</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>JJA</td>
<td>DJF</td>
<td>JJA</td>
<td>DJF</td>
<td>JJA</td>
</tr>
<tr>
<td>N</td>
<td>10.2</td>
<td>10.0</td>
<td>8.5</td>
<td>8.8</td>
<td>10.7</td>
</tr>
<tr>
<td>OLR</td>
<td>6.5</td>
<td>6.3</td>
<td>8.4</td>
<td>8.8</td>
<td>4.8</td>
</tr>
<tr>
<td>Albedo</td>
<td>1.2</td>
<td>1.1</td>
<td>4.1</td>
<td>4.2</td>
<td>1.1</td>
</tr>
<tr>
<td>Net</td>
<td>2.4</td>
<td>2.3</td>
<td>-6.4</td>
<td>-7.5</td>
<td>1.4</td>
</tr>
</tbody>
</table>


to longwave and shortwave forcing at the TOA, a study extended to the atmospheric and surface radiation budgets by Chen et al. (2000). They used a simplified group of clouds (5 instead of 9) from the ISSCP database. The results of their study are shown in Table 10.6. Their analysis demonstrated the dominance of high clouds on the outgoing longwave radiation (OLR), at least in low latitudes, as well as the important role of thicker, higher clouds in the OLR of the Tropics and low clouds on the shortwave fluxes in the mid- and high-latitudes.

Cirrus clouds exert varying but significant effects on the radiation budget of the earth. When the optical depth of cirrus is sufficiently low, the absorption and re-emission of IR dominates the solar albedo effect and cirrus clouds then exert a warming effect. As the optical depth increases, the albedo effect increases which leads to a net cooling. In terms of the surface and atmospheric radiation budgets, thin cirrus cool the surface, but exert a net warming within and at the top of the atmosphere. Optically thick cirrus, on the other hand, still warm the atmosphere on the whole, but cool the surface and the upper atmosphere (Chen et al., 2000). In spite of their importance in the radiation budget, the understanding of the global net radiative responses to the various cirrus cloud types is still not well known.

As stated above, the altitude and vertical extent of clouds are important parameters in the radiation balance of the earth. A cirrus cloud at high altitudes will exert a more significant impact on the IR flux than the same cirrus cloud at lower altitudes. A cirrus cloud at lower altitudes may exert less of an impact on the IR radiative flux, and hence the albedo effect may become dominant. For a given cloud, the cloud may transition from cooling the surface to warming the surface relative to clear-sky conditions if the surface albedo is increased. The radiative effects of cirrus depend strongly on their microphysical properties in contrast to those of low and middle clouds. For example, for the same optical thickness or IWP, the forcing associated with cirrus clouds can switch
sign depending on the shape and size of the ice crystals (Zhang et al., 1999). Unfortunately, as we saw above, the microphysical properties of cirrus are still not well known.

Other factors that influence the radiative effect of cirrus clouds include cloud fraction, variations in horizontal homogeneity, the solar zenith angle, the ice crystal size distribution, the cloud geometrical thickness, IWC and the presence of other clouds or water vapor in the column. Schlimme et al. (2005) noted from their sensitivity study that the order of importance of cirrus cloud properties in solar broadband radiative transfer is optical thickness, ice crystal shape, ice particle size and spatial structure. Cirrus clouds also affect the radiation budget indirectly through their influence on water vapor distribution in the Tropics. In regions of subsidence, away from deep convective outflow, cirrus clouds tend to moisten the upper troposphere, which in turn influences the heating rate.

The optical depths of in situ cirrus are not often greater than 5, while the optical depths of anvil cirrus can range from 10-50 and those for contrails are of the order of 0.1-0.5. Solar radiative effects appear to be more important for optical depths in the range of 1-2, whereas IR effects are more important for thicker layers in which the optical depth is greater than 2. For optical depths of less than 1, radiative effects appear minimal. Subvisible cirrus layers can have heating rates of up to 1 K day$^{-1}$ and radiative forcing of 1.2 W m$^{-2}$, which is equivalent to ~ 0.7 °C change in surface temperature globally (McFarquhar et al., 2000). Haladay and Stephens (2009) estimated that the effects of their thin ice cloud category was ~ 4 W m$^{-2}$ on the average IR heating of the tropics. Radiative forcing can increase both the IWP and cirrus lifetime through dynamic responses, although this is dependent on the cloud layer stability. Anvil cirrus or in situ mid-latitude cirrus can be enhanced through turbulence effects forced by radiative heating in the lower regions of the cloud. Radiative cooling at cloud top may also produce supersaturations that are sufficiently high to initiate ice nucleation.

The effects of radiative heating on convection have been demonstrated to regulate high level cloudiness, such as in the global study by Fowler and Randall (1994). More recently, Stephens et al. (2008) examined the impacts of radiative heating on convection using a cloud-resolving model in radiative convective equilibrium. It was evident from their model sensitivity experiments, one in which the radiative forcing was held constant irrespective of the cloud amounts, and the other in which the optical properties of clouds were turned off, that the lack of upper-tropospheric radiative heating associated with these anvil clouds influences the stability of the atmosphere in such a manner that convection was strengthened, thus producing more high clouds relative to the control experiment in which the radiative forcing and optical properties were not altered. The high clouds were found to increase by a factor of 6-8 over the amount of high clouds produced in the control simulation. When the radiative heating of these high clouds was included, as in the control simulation, the high cloud fraction was significantly reduced, even though the actual fractional area
of convection was not significantly altered. Thus the high cloud radiative heating appears to provide a feedback on the convection that regulates the high clouds produced by the convection.

### 10.4.6. Aerosol Effects on Middle- and High-Level Clouds

The relatively high number concentrations of small ice crystals (∼0.1-10 cc⁻¹) regularly observed in in situ cirrus clouds in both the tropics and mid-latitudes appear to be associated with the mesoscale temperature variations generated by the variations in vertical velocity within cirrus clouds (Kärcher and Ström, 2003; Jensen and Pfister, 2004; Hoyle et al., 2005). Observations of the tropical and mid-latitude upper troposphere suggest a relatively consistent distribution of mesoscale temperature fluctuations, leading to typical cooling rates of the order of 10 K hr⁻¹. The origin of these fluctuations is not well known although they appear to be associated with gravity waves. In such a dynamic environment, the effect of an IN population on cirrus properties depends primarily on the IN number concentration, the ice nucleation threshold (determined by size, chemical composition etc.) and the local cooling rate. As described in greater detail below (references included there), it has been hypothesized that the likely impact of enhanced IN concentrations, keeping all other factors constant, is to reduce homogeneous freezing and decrease the total number of ice crystals formed, thus producing slightly larger effective ice crystal sizes and less bright cirrus. Absorption of thermal emission by large ice crystals may induce internal convective instability (especially in less stable thermodynamic cloud environments), thereby prolonging cirrus lifetimes by additional cooling and possible triggering of turbulence-induced ice nucleation. Unfortunately the response of these processes to enhanced IN concentrations have not as yet been observed in the field. Also, our current inability to accurately measure RHI in the upper troposphere limits our ability to discriminate between the various ice nucleation processes associated with different IN types.

While chemical composition and size are important considerations for a given ice nucleus, the general activation of ice crystals appears to be less sensitive to the size distribution of freezing aerosol and more sensitive to the updraft speeds and associated cooling rates. Homogeneous and heterogeneous nucleation both contribute to the formation of ice in middle and high-level clouds. Secondary ice processes also play a role in middle-level clouds but appear to be less important in cirrus clouds (Cantrell and Heymsfield, 2005). For cirrus clouds, the predominant heterogeneous modes are immersion freezing and deposition nucleation (DeMott, 2002). While immersion nuclei are likely to be observed in the upper troposphere, deposition nuclei often also act as CCN and thus may be removed before reaching the upper troposphere. IN may also be pre-activated in that following sublimation, the IN may later generate ice at lower RHI compared to previously unactivated IN. Twohy et al. (2009) demonstrated that Saharan dust particles can serve both as CCN and
IN. This has important implications for mixed-phase clouds and anvil cirrus as Saharan and Asian dusts activate more efficiently as condensation freezing nuclei than as deposition nuclei at temperatures warmer than \( \sim 240 \) K (Field et al., 2006). The contributions of the homogeneous freezing of liquid particles and heterogeneous ice nucleation involving mixed-phase or insoluble particles vary for cirrus that form in situ. While direct ice nucleation from aerosol particles may contribute during the life cycle of anvil cirrus and contrail cirrus, this does not appear to be the primary factor in in situ cirrus. Heymsfield et al. (1991) note that, as IN concentrations are low in cirrus clouds and yet ice crystals are evident in these clouds, ice crystal generation below \(-40 \, ^\circ\mathrm{C}\) is due primarily to the homogeneous freezing of water droplets. Within jet exhaust, high concentrations of aerosols are activated \( \left(10^4 - 10^5 \, \text{cc}^{-1}\right)\) into water droplets that freeze rapidly. The actual nucleation process appears to be less important in contrails compared to the high number of available aerosols and the rapid cooling rates within these systems. Ice processes are discussed in more detail in Chapter 4.

Aerosol indirect effects associated with IN on middle- and high-level clouds are still not well understood. Changes in the number concentrations and sizes of liquid water particles can produce changes in the total concentrations of ice crystals of the order of a factor of 2 (Kärcher and Lohmann, 2002). When efficient IN and liquid particles compete during the formation of cirrus clouds, aerosol indirect effects may become more significant (DeMott et al., 1997). For a given cooling rate in a rising air parcel, the activation and deposition of water vapor on IN-generated ice crystals at low supersaturations reduces any increases in RHI that would otherwise occur, thus producing fewer homogeneously frozen particles. Adding IN to a population of liquid particles can therefore lead to a reduction in the number of nucleated ice crystals (DeMott et al., 1994). This process has been called the **negative Twomey effect** (Kärcher and Lohmann, 2003) in that while the Twomey effect refers to an increase in the number of activated cloud droplets with increasing CCN concentrations, this effect instead reduces the number of activated ice crystals with enhanced IN concentrations (DeMott et al., 1994). The negative Twomey effect can reduce total ice crystal concentrations by a factor of 10. Model simulations have demonstrated that it is associated with increased effective radii and reduced IWCs, and hence reduced cirrus albedos, as well as changes in optical extinction, subvisible cloud fraction and the occurrence of cirrus (Haag and Kärcher, 2004).

The magnitude of IN aerosol indirect effects depends on the ratio between the cooling rate and the growth rate. If sufficient IN are available for activation and the cooling rates are relatively slow so that only a fraction of these IN are activated, more ice crystals could then form than in the case when no IN are available. However, this effect tends to be weaker than the negative Twomey effect (Kärcher et al., 2006). In situ measurements of heterogeneous nucleation on IN have shown that the presence of IN did not prevent homogeneous freezing from occurring (DeMott et al., 2003), and while the negative Twomey effect has
been inferred from satellite data, lidar observations of the same case study were inconclusive (Seifert et al., 2007). It is thus very difficult to directly attribute differences in cirrus properties to variations in IN concentrations given the difficulties in separating the dynamic effects from the aerosol forcing. More recently the negative Twomey effect has been studied in large-eddy simulations on the cloud scale (Spichtinger and Gierens, 2008). These simulations show that under some conditions the macroscale evolution of cirrus may be strongly affected by IN, while overall radiative properties are still dominated by liquid particle freezing. It is thus clear that significant advances are required in our understanding of aerosol indirect effects on cirrus.

While convective storms potentially transport IN from the surface to the upper troposphere it is uncertain as to what fractions of the IN actually reach the anvil cirrus and as to how significant their impacts may be. Very few observational studies have been conducted of the impacts of aerosols on deep convection and their associated anvils. Most of what is understood in this regard has been obtained via numerical modeling. Two recent modeling studies investigated the impacts of CCN and IN on the characteristics of deep convection and their associated anvils observed during the CRYSTAL-FACE field campaign (van den Heever et al., 2006; Carrió et al., 2007). Using a cloud resolving model, van den Heever et al. (2006) demonstrated that, in the presence of enhanced CCN and IN concentrations associated with a Saharan dust intrusion, the anvils covered a smaller area but were better organized and had larger condensate mixing ratios. Higher IN concentrations produced ice at warmer temperatures (and hence lower in the atmosphere) and the anvils were found to be thicker. Based on the results of their modeling study of storms that developed during the same field campaign, Fridlind et al. (2004) advanced the hypothesis that aerosols between 6 and 10 km had the greatest impact on the cirrus anvil microphysics. However, the simulations presented by van den Heever et al. demonstrated that many characteristics of both the convective and anvil stages of storm development were more sensitive to changes in the aerosol concentrations below 4 km. Other modeling studies have shown greater number concentrations of smaller crystals in response to enhanced aerosol concentrations (Khain et al., 2004) while others have observed the opposite trend, which they have attributed to differences in the strength of convection (Cui et al., 2006). Still other modeling studies have emphasized the role of aerosols on the release of latent heat and the feedbacks to subsequent storms dynamics (e.g. Khain et al., 2005; van den Heever et al., 2006).

In a subsequent study, Carrió et al. (2007) made use of an LES model forced by the output data from the mesoscale simulations of van den Heever et al. (2006), to investigate the anvil processes and characteristics in detail. The results from the LES simulations showed that variations in CCN and IN have significant effects on the optical properties and lifetime of anvil cirrus clouds. The anvils were both physically and optically thicker under enhanced aerosol concentrations. The largest differences in optical depth (∼40%), cloud-averaged
effective diameter (>25%) and IWC (~75%) occurred during the first hours of the simulation due to a higher frequency of large particles (aggregates). These differences then decreased due to the sedimentation of the aggregates. Toward the end of the simulations, the cloud-averaged effective diameter showed little change compared with the clean case. However, given that the differences in ice particle number concentrations remained large (>30%) throughout the simulation, the impacts of nucleating aerosol on optical depth continued to be significant. All of the microphysical and optical properties demonstrated a monotonic response to enhanced nucleating aerosol concentrations. Whether increasing those aerosols that serve as both CCN and IN, or just those serving as CCN, the mass fraction of particles involved in collisions (aggregates) and the total ice number concentrations increased monotonically, while those ice crystals that grew only by vapor diffusion (pristine ice) showed no significant changes. On the other hand, increasing the concentrations of those aerosols that serve as IN had a significant effect on the mass fraction of pristine ice but little effect on aggregates. The largest differences in the size distributions is associated with those simulations in which aerosols may serve as both CCN and IN. Enhancing CCN had two different effects on the simulated particle size distributions: (1) a large number of smaller, supercooled cloud droplets reached the top of the convective cloud, thereby significantly increasing the number concentration of ice particles with diameters less than 20 microns; and (2) significantly larger numbers of cloud droplets at high levels of the convective cloud increased the probability of binary interactions between particles and therefore the production of aggregates. It appears that the influence of enhanced concentrations available to act as CCN exerted the dominant effect on the particle size distributions of anvil cirrus. Finally, the narrower ice particle size distributions and the lower turbulent kinetic energy associated with enhanced aerosol number concentrations resulted in enhanced anvil-cirrus lifetimes.

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