(Fig. 1.26). A sketch illustrating the factors involved in formation of the banner cloud is shown in Fig. 12.23. As indicated, a very sharp-edged peak is usually involved in the formation of a banner cloud. When the flow encounters the peak, it is both forced up over the peak and split laterally into two branches. The separation of the flow in the vertical and horizontal leads to an increase of speed and corresponding reduction of the pressure [according to (8.42)] at the downwind edge of the mountain. A minimum of pressure perturbation is thus created in the lee of the top part of the peak. The pressure gradient force thus created generates [according to (7.1)] vertical motion up the lee slope. If the air drawn up the lee slope is sufficiently moist and reaches its condensation level below the top of the mountain, cloud forms. When the newly formed cloud reaches the height of the top of the mountain, it is incorporated into the main airstream over the mountain and is swept rapidly downwind in a plume resembling a wind sock or smoke from a chimney.[332]

12.4 Orographic Precipitation

So far we have discussed nonprecipitating clouds that form in shallow moist layers in flows perturbed as they flow over ridges or peaks in the surface topography. It is also possible for orographically induced flow to produce or influence precipitating clouds. Figure 12.24 summarizes the mechanisms of orographic control over precipitation. We will consider each of these mechanisms briefly in the following subsections.

12.4.1 Seeder–Feeder Mechanism over Small Hills

In Sec. 6.2, we saw how the stratiform precipitation process can be enhanced by the seeder–feeder mechanism, in which convective cells aloft can produce large precipitation particles, which, upon falling through a lower cloud layer, grow at the expense of the water content of the lower cloud. Stratus or small cumulus formed in the boundary-layer flow over small hills (Sec. 12.1) can be a particularly effective feeder cloud. As illustrated schematically in Fig. 12.24a, precipitation from another cloud layer aloft may be enhanced as it falls through the low-level feeder cloud. By itself, the low-level cloud might not precipitate. Precipitation particles from the upper cloud collect cloud particles from the low cloud, and the water thus collected is deposited on the ground.[333] Figures 12.25 and 12.26 illus-

[332] This plausible explanation of the banner cloud appeared in the extensive nineteenth-century treatment of the global system, *Die Erde als Ganzes* (The Earth as a Whole) *ihre Atmosphäre und Hydrosphäre*, by Hann (1896). An essentially similar explanation was offered by Douglas (1928). He, however, neglected to mention the possible role of horizontal flow separation. Unfortunately, a variety of specious explanations of the banner cloud have been suggested and have permeated the literature on the subject since these early studies.

[333] This mechanism was proposed by Bergeron (1950, 1968) to explain the enhancement of precipitation over low hills. As we have seen in Chapter 6, the idea not only was accepted for orographic enhancement but is now thought to be important in stratiform precipitation in general.
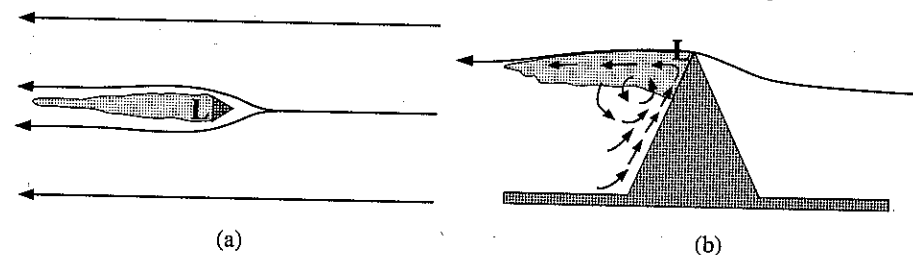
(a)   (b)

Figure 12.23 Schematic illustration of the formation of a banner cloud by flow past an isolated mountain peak. (a) Plan view (b) vertical cross section. Light stippling denotes the cloud.
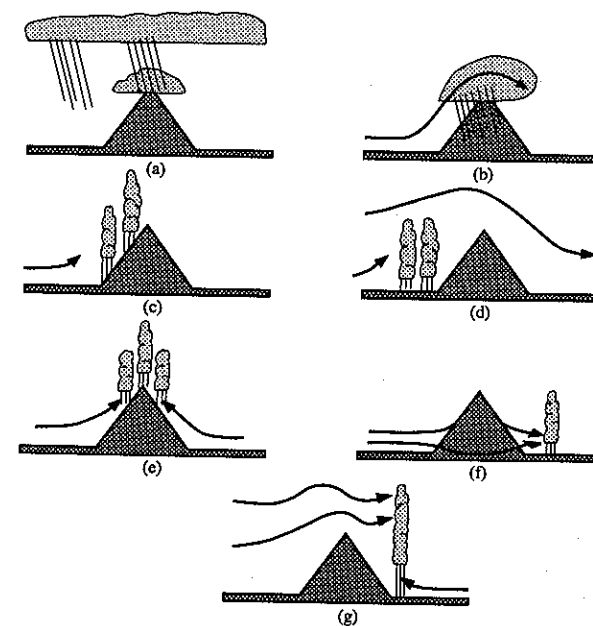

Figure 12.24 Mechanisms of orographic precipitation. (a) Seeder–feeder mechanism; (b) upslope condensation; (c) upslope triggering of convection; (d) upstream triggering of convection; (e) thermal triggering of convection; (f) lee-side triggering of convection; (g) lee-side enhancement of convection. Clouds are lightly stippled. Slanted lines below cloud base indicate precipitation.

trate an example of this process over low hills in South Wales. In this case, the upper-level clouds are moving, orographically triggered convective clouds. However, the process works equally well if the pre-existing clouds are of some other type (e.g., frontal).

12.4.2 Upslope Condensation

As we have seen, it is possible for stable ascent forced by flow over a mountain ridge or peak to be felt through a deep layer above the mountain. If the air forced over a mountain is sufficiently moist through a large portion of the lifted layer,
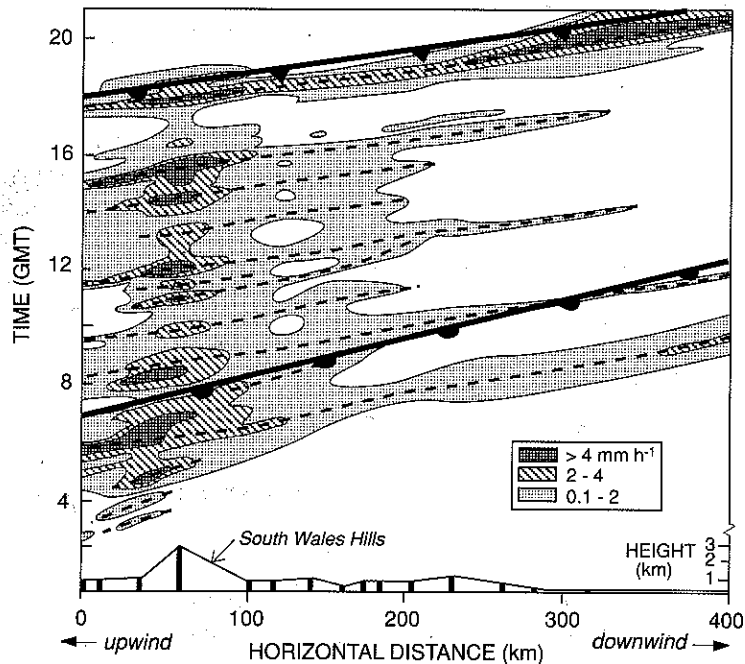
Figure 12.25 Time–distance diagram showing the rainfall rates (mm h^{-1}) and the movement of mesoscale precipitation areas over the hills of South Wales. The rainfall begins ahead of the cold front and continues in the warm sector. The rainfall rate over the hills is continuous but variable and closely associated with the passage of convective clouds aloft. (Adapted from Smith, 1979, after Browning et al., 1974.)

condensation may occur through a deep layer [in contrast to shallow moist layers leading to boundary-layer clouds in upslope flow (Sec. 12.1) or wave clouds (Secs. 12.2 and 12.3)]. Figure 12.24b illustrates this process conceptually. Figure 12.27 illustrates the amount of cloud water that can be produced by flow over a mountain ridge. The streamlines, calculated with a nonlinear two-dimensional model (Fig. 12.27a), represent flow over the Cascade Mountain Range of Washington State under typical wintertime conditions. The isopleths of cloud liquid water content calculated by means of the bulk nonprecipitating warm-cloud water-continuity scheme [(3.63)–(3.64)] are shown in Fig. 12.27b. The maximum water content exceeds 1 g kg^{-1}, which is large enough to expect precipitation to develop [recall that the autoconversion threshold in (3.76) is often assumed to be 1 g kg^{-1}].

It is possible for precipitating clouds to occur in pure orographic flow of the type indicated in Fig. 12.27a. More often, though, the orographically generated precipitating cloud is superimposed on pre-existing clouds, especially those associated with a front passing over the mountain range, in which case the water field in Fig. 12.27b can be regarded as the potential enhancement of the frontal precipitation by the orographic upward motion. The downward motion on the lee side, of course, dries out the frontal clouds after they pass over the ridge.
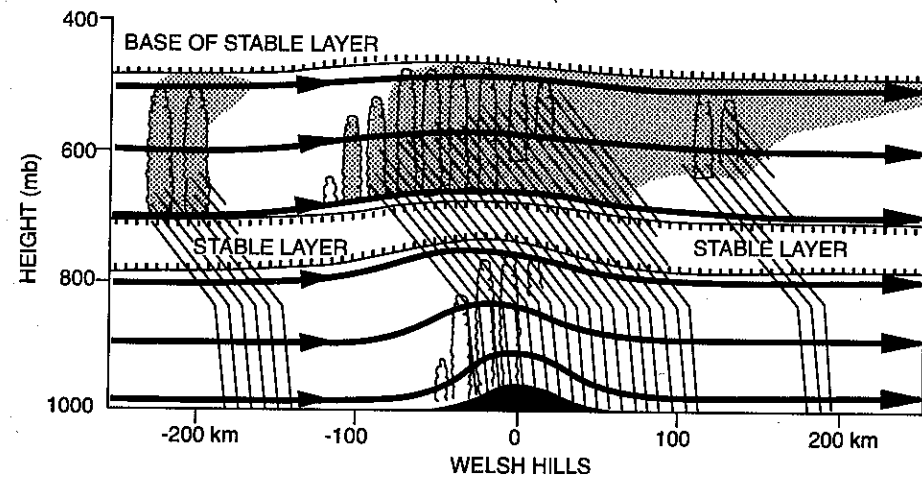

Figure 12.26 Schematic cross section of the rain clouds in a warm sector over hills of South Wales. The moving, orographically triggered convective clouds aloft produce precipitation which is locally enhanced by the low-level convective clouds over the hills (see Fig. 12.25). The slope of the hydrometeor trajectories changes abruptly at the freezing line as snow changes to rain. (Adapted from Smith, 1979, after Browning et al., 1974.)
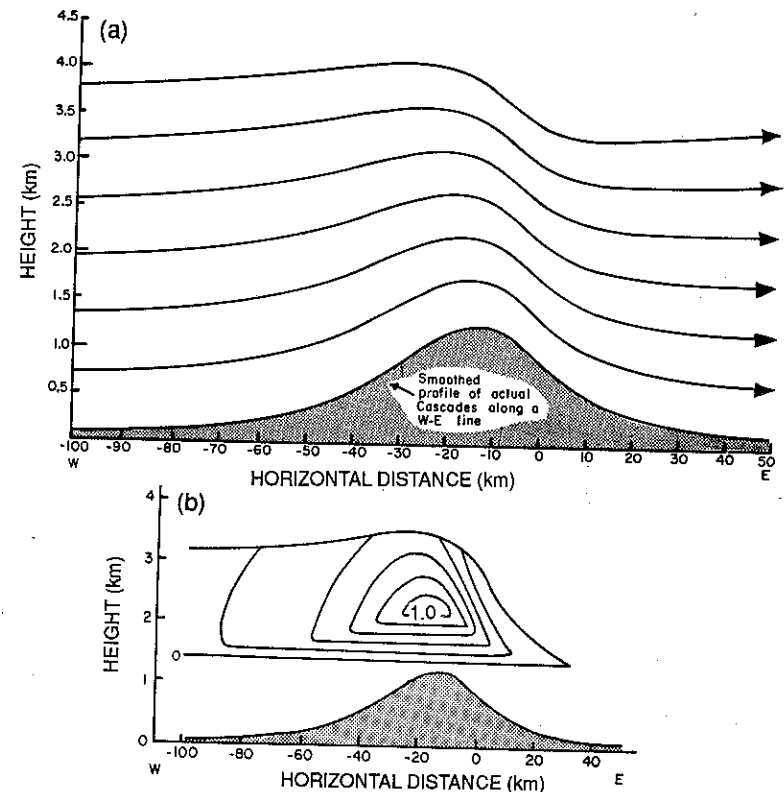

Figure 12.27 Model calculation of condensation occurring in stable ascent forced by a typical westerly flow over the Cascade Mountains of Washington State. (a) Streamlines of the flow. (b) Isopleths of adiabatic condensate at 0.2 g kg^{-1} intervals. (Panel a from Fraser et al., 1973; panel b from Hobbs et al., 1973. Reprinted with permission from the American Meteorological Society.)

The situation depicted in Fig. 12.27 is highly idealized in that real mountain ranges are not perfectly two-dimensional. We saw in Sec. 12.3 that when $\bar{u}/N\hat{h}_m$ is small, the flow tends to turn laterally and go horizontally around peaks rather than over them. It is indeed often the case that $\bar{u}/N\hat{h}_m$ is small, such that two-dimensional flow is blocked and three-dimensional flow around peaks and through gaps becomes significant, and vertical motions are correspondingly reduced.

12.4.3   Orographic Convection

When the air flowing over rugged terrain is potentially unstable, the lifting induced by the terrain can lead to the release of instability. In this case, the orographic clouds take the form of cumulus or cumulonimbus rather than fog, stratus, wave clouds, or stable precipitating clouds. Orographic cumulonimbi can be very important precipitation producers. Some of the rainiest areas of the world (e.g., the monsoon areas of India, Bangladesh, and Southeast Asia) are dominated by this type of precipitation.

Once formed and active, the orographic convective clouds are largely governed by the dynamics of convective clouds, as discussed in Chapters 7–9. As long as the clouds remain in the vicinity of the mountain, however, the cloud dynamics will be a complex interaction of convective and orographic dynamics. These complex interactions are only beginning to be understood. We will therefore not delve too deeply into this topic. We will concern ourselves only with the triggering and enhancement that occur during the earlier stages of the orographic convection. These processes are indicated in Fig. 12.24c–g, which indicate how various flows over and around topography can trigger or enhance convection upstream of the mountain, on the windward slope, directly over the peak, on the lee slope, or downwind of the mountain.

12.4.3.1   Upslope and Upstream Triggering

It is fairly obvious that any upslope motion can trigger convection if the air moving upslope is sufficiently moist and unstable. Figure 12.24c represents such upslope triggering. It is perhaps less obvious that topographically induced motions may lead to condensation and triggering of convection upstream of the mountain slope. However, as we have seen, flow over mountains can become complex above the surface layer, with lifting induced by the terrain sometimes being felt aloft some considerable distance upstream of the mountain. The idea of upstream triggering is indicated conceptually in Fig. 12.24d and further illustrated in Fig. 12.28, where the layer of air between the ground and the streamline is shown to be destabilized as the parcel upstream of the hill is lifted while going from A to B.

Two types of upstream lifting may occur. We have seen that vertically propagating waves described by linear theory, which are associated with mountains of any size, can tilt upstream (Figs. 12.3 and 12.4). The case shown in Fig. 12.26 illustrates how convective cells can be triggered upstream of small hills. Such a situation may be difficult to recognize in practice since as the clouds form and develop they are advected over the hills.

A second type of upstream lifting is associated with blocking (i.e., with nonlinear effects introduced by a large, finite mountain barrier). An intuitive feeling for how blocking by a two-dimensional obstacle can produce upstream lifting is provided by examining what happens in a controlled laboratory setting when a homogeneous layer of fluid flowing in a restricted channel at velocity $\bar{u}$ with low Fr is blocked by a barrier. Figure 12.29 idealizes this situation for two cases. In the case of complete blocking, the Fr is too low for the fluid to surmount the barrier; consequently, the fluid piles up against it. Mass continuity is satisfied by a hydraulic jump upstream (at x_j). Since mass continuity must be satisfied across the jump, then, according to (12.28), u decreases with increasing x across the jump. Thus, either the fluid parcels at x_j are rising as the jump travels out from the barrier, or the depth of the fluid to the right of x_j is increasing as mass piles up against the barrier. In either case, upward motion could be occurring at $x \geq x_j$. However, neither of these two situations is steady state.

A steady state can be achieved if the flow is partially blocked, as shown in Fig. 12.29b. In a laboratory setting, this situation is produced by towing the obstacle along the bottom of a tank. This case corresponds to the flow associated with severe downslope winds and rotor cloud formation on the lee side (Sec. 12.2.5). The partial blocking phenomenon is the windward-side portion of the flow. This case can be in a steady state since the blocking is incomplete and downstream and upstream conditions can now be matched. The upstream jump thus can become stationary in a frame moving with the laboratory obstacle, and a fixed region of upward motion can be established upstream of the obstacle. It is presumed that an analogous situation can occur in the atmosphere.

One area where upstream partial blocking is thought to play a role in triggering cumulonimbus is over the Arabian Sea west of the Western Ghats Mountain Range of southwestern India. Figure 12.30 contains the results of a nonlinear, two-dimensional model simulation (using a model of the type discussed in Sec. 7.5.3) of the atmospheric flow over the Western Ghats. The upstream flow was taken to be that characteristic of the Indian summer monsoon. The mountain profile is a smoothed representation of the Western Ghats, which slope up steeply from the ocean and level out into a plateau. The pressure perturbation field (Fig. 12.30a) indicates the piling up of mass against the windward slope. The horizontal wind field (Fig. 12.30b) shows a corresponding deceleration as the lower-tropospheric monsoon flow approaches the range. Finally, the vertical velocity field (Fig. 12.30c) shows how the lifting extends out over the ocean, where it could trigger convection upstream of the mountain range. It should be noted, however, that once convection is initiated, the flow can be altered by the convective heating. Thus, the situation portrayed in Fig. 12.30 may not be the final state of affairs.

12.4.3.2   Thermal and Lee-Side Triggering

Figure 12.24e and f indicate two additional ways in which mountains can trigger convection. Thermal forcing (Fig. 12.24e) occurs when daytime heating produces an elevated heat source and corresponding thermally direct circulation, with con-
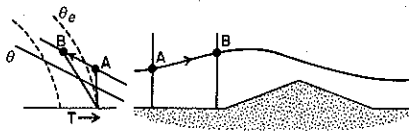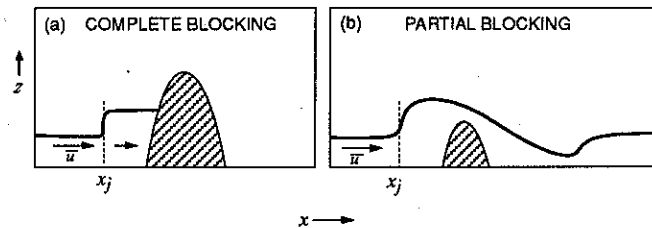
Figure 12.28

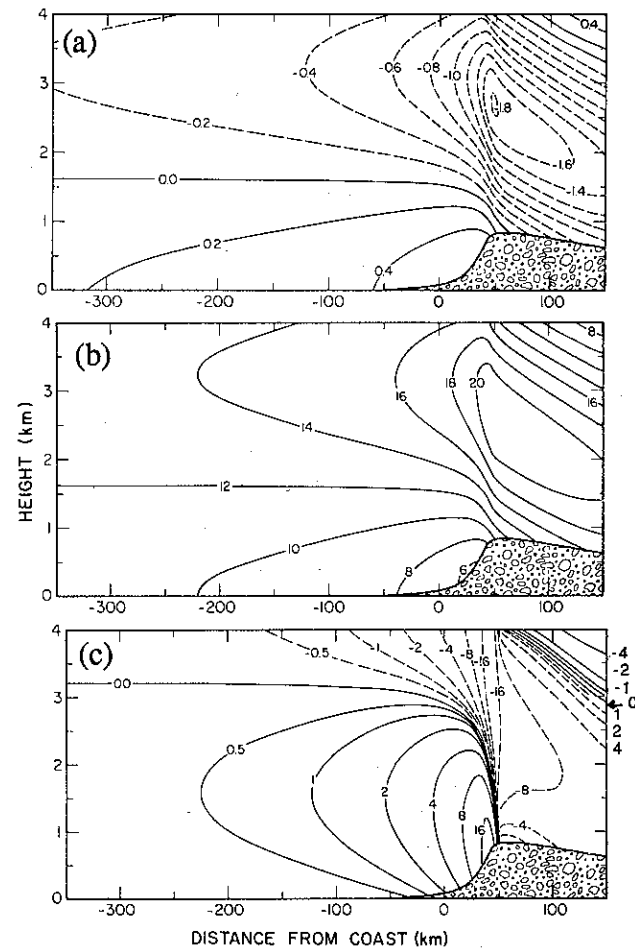**Figure 12.29**

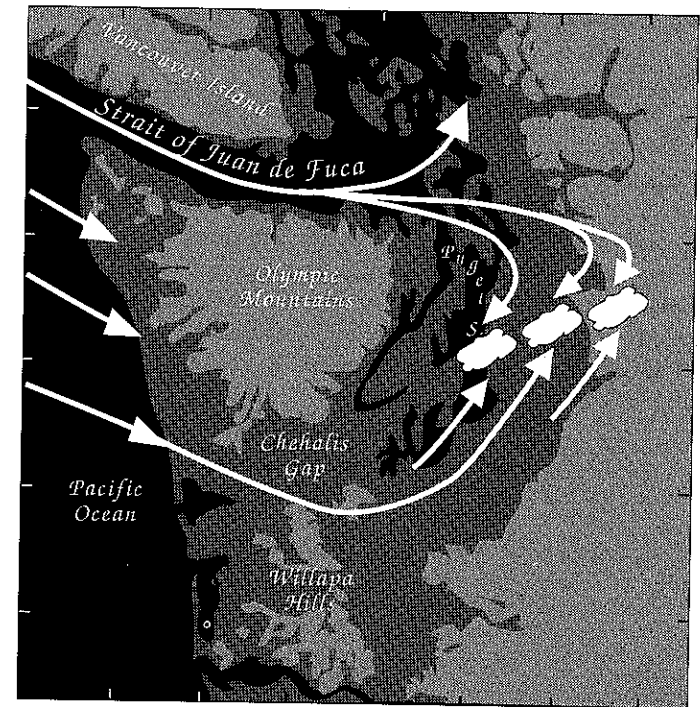DISTANCE FROM COAST (km)

Figure 12.30

Figure 12.31   Triggering of convection in the lee of the Olympic Mountains under conditions of the Puget Sound Convergence Zone of Washington State. Streamlines represent surface winds. (Adapted from Mass, 1981. Reproduced with permission from the American Meteorological Society.)

vergence at the top of the mountain. It is well known that this type of circulation can trigger anything from small cumulus to incipient mesoscale convective systems, all of whose dynamics have been discussed in previous chapters.

The lee-side forcing indicated in Fig. 12.24f results from low-Froude-number flow around an isolated obstacle. One consequence of this diversion around the mountain is the formation of convection in the lee of the mountain. A good

Figure 12.28   Destabilization of a nearly saturated, nearly conditionally unstable air mass owing to lifting aloft upstream of the mountain. The air parcels aloft rise dry adiabatically, thus decreasing the stability of the air column against moist convection. (From Smith, 1979.)

Figure 12.29   Blocking of a homogeneous layer of fluid flowing in a restricted channel at a low Froude number. (a) Complete blocking. (b) Partial blocking. Stationary obstacle shown by hatching. Fluid enters from left at velocity $\bar{u}$. (Adapted from Simpson, 1987. Reproduced with permission from Ellis Horwood Ltd., Chichester.)

Figure 12.30   Nonlinear two-dimensional model simulation illustrating upstream partial blocking over the Arabian Sea west of the Western Ghats Mountain Range of southwestern India. Cross sections perpendicular to mountains of (a) pressure perturbation (mb); (b) horizontal wind speed (m s^{-1}); and (c) vertical velocity (cm s^{-1}). (From Grossman and Durran, 1984. Reprinted with permission from the American Meteorological Society.)

example of this phenomenon is the Puget Sound Convergence Zone, which forms and triggers convection in the lee of the Olympic Mountains of Washington State (Fig. 12.31).

12.4.3.3 Lee-Side Enhancement of Deep Convection

The sketch in Fig. 12.24g indicates how convection triggered on the windward slope over the crest of a ridge can be enhanced on the lee side. The enhancement results from combined effects of midlevel upward motion associated with a vertically propagating wave induced by flow over a mountain and low-level thermally induced upslope flow. This combination of effects is thought to be significant in the enhancement of deep convection that forms over the Rocky Mountains and subsequently develops into mesoscale convective complexes that move eastward across the central United States.[334]

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[334] See Tripoli and Cotton (1989a) for a detailed conceptual model of this phenomenon.

References

Abe, M., 1932: The formation of cloud by the obstruction of Mount Fuji. *Geophys. Mag.*, **6**, 1–10.

Abe, M., 1941: Mountain clouds, their forms and connected air current. Part II. *Bull. Central Met. Obs. Japan*, VII, **3**, 93–145.

Acheson, D. J., 1990: *Elementary Fluid Dynamics*. Clarendon Press, Oxford, 397 pp.

Ackerman, T. P., K.-N. Liou, F. P. J. Valero, and L. Pfister, 1988: Heating rates in tropical anvils. *J. Atmos. Sci.*, **45**, 1606–1623.

Agee, E. M., 1982: An introduction to shallow convective systems. *Cloud Dynamics* (E. M. Agee and T. Asai, Eds.), D. Reidel Publishing, Dordrecht, 423 pp.

Agee, E. M., T. S. Chen, and K. E. Doswell, 1973: A review of mesoscale cellular convection. *Bull. Amer. Meteor. Soc.*, **10**, 1004–1012.

Anderson, R. K., J. P. Ashman, G. R. Farr, E. W. Ferguson, G. N. Isayeva, V. J. Oliver, F. C. Parmenter, T. P. Popova, R. W. Skidmore, A. H. Smith, and N.F. Veltishchev, 1973: *The Use of Satellite Pictures in Weather Analysis and Forecasting*. Technical Note No. 124, World Meteorological Organization, Geneva, 275 pp.

Anderson, R. K., et al., 1974: Application of Meteorological Satellite Data in Analysis and Forecasting. ESSA Technical Report NESC 51, U.S. Dept. of Commerce, NTIS AD-786 137.

Anthes, R. A., 1982: *Tropical Cyclones: Their Evolution, Structure and Effects*. American Meteorological Society, Boston, 208 pp.

Árnason, G., and R. S. Greenfield, 1972: Micro- and macro-structures of numerically simulated convective clouds. *J. Atmos. Sci.*, **29**, 342–367.

Arya, S. P., 1988: *Introduction to Micrometeorology*. Academic Press, New York, 303 pp.

Asai, T., 1964: Cumulus convection in the atmosphere with vertical wind shear: Numerical experiment. *J. Met. Soc. Japan*, **42**, 245–259.

Asai, T., 1970a: Three-dimensional features of thermal convection in a plane couette flow. *J. Met. Soc. Japan*, **48**, 18–29.

Asai, T., 1970b: Stability of a plane parallel flow with variable vertical shear and unstable stratification. *J. Met. Soc. Japan*, **48**, 129–139.

Asai, T., 1972: Thermal instability of a shear flow turning direction with height. *J. Met. Soc. Japan*, **50**, 525–532.

Asai, T., and A. Kasahara, 1967: A theoretical study of the compensating downward motions associated with cumulus clouds. *J. Atmos. Sci.*, **24**, 487–496.

Asai, T., and I. Nakasuji, 1973: On the stability of the Ekman boundary-layer flow with thermally unstable stratification. *J. Met. Soc. Japan*, **51**, 29–42.

Atlas, D., Ed., 1990: *Radar in Meteorology*. American Meteorological Society, Boston, 806 pp.

Atlas, D., K. R. Hardy, R. Wexler, and R. J. Boucher, 1963: On the origin of hurricane spiral bands. *Technical Conference on Hurricanes and Tropical Meteorology*, Mexico, *Geofisica Internacional*, **3**, 123–132.

Atlas, D., D. Rosenfeld, and D. Wolff, 1990: Climatologically tuned reflectivity rain rate relations and links to area–time integrals. *J. Appl. Meteor.*, **29**, 1120–1135.

Atwater, M. A., 1972: Thermal effects of organization and industrialization in the boundary layer: A numerical study. *Boundary-Layer Met.*, **3**, 229–245.

Auer, A. H., Jr., 1972: Distribution of graupel and hail with size. *Mon. Wea. Rev.*, **100**, 325–328.