Applied Sci & Tech PhD Program, North Carolina A&T State University EES 885-3

Dynamical Mountain Meteorology

Dr. Yuh-Lang Lin, ylin@cat.edu; http://mesolab.org
North Carolina A&T State University
(Ref.: *Mesoscale Dynamics*, Y.-L. Lin, Cambridge, 2007)

Chapter 12 Other Orographic Effects

(Based on Sec. 5.6 of Lin 2007)

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In addition to the phenomena discussed in previous sections, orography also plays a significant role in influencing the general airflow and weather systems, such as frontal passage, the formation of coastally trapped disturbance, cold-air damming, gap flow, orographic precipitation and foehn wind.

The dynamics of orographic precipitation and foehn wind require backgrounds in thermally forced flow, mesoscale instabilities, and moist convection. Descriptions of these are made in Chapter 11. The orographic effects on other phenomena are briefly described next.

12.1 Effects on Frontal Passage

When a front passes over a mesoscale mountain range, it is often distorted by the mountain. As soon as the front impinges on the mountain, an anticyclonic circulation, associated with the mountain-induced high pressure, is produced in the cold postfrontal air and the front advances faster on the left (facing the frontal movement).

Figure 5.38 shows a conceptual model of a two-dimensional front passing over a mesoscale mountain ridge.

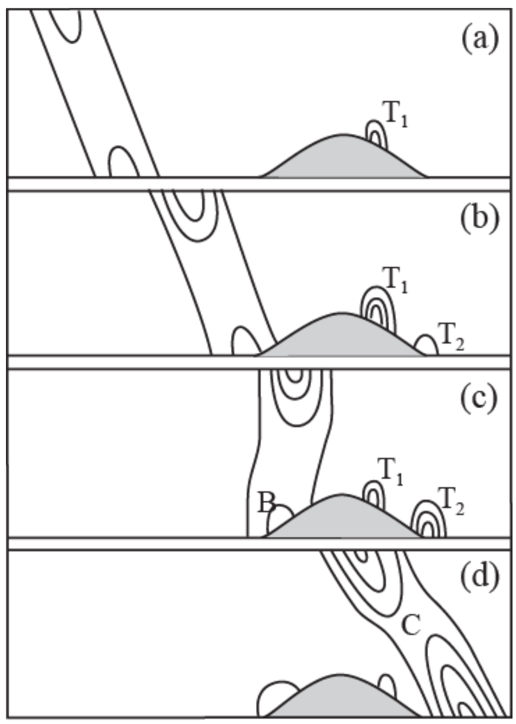

Fig. 5.38: A conceptual model of a two-dimensional frontal passage over a mesoscale mountain, which indicates the low-level blocking (denoted by B) of the front along the windward slope, the development of the lee trough (T_1) and secondary trough (T_2) along the lee slope, the separation of the upper-level and lower-level frontal waves, and the coupling (C) of the upper-level frontal wave with the secondary trough in the lee of the mountain. (Adapted after Dickinson and Knight 1999)

- The mountain blocks and decelerates the approaching front at the surface while the upper-level potential vorticity anomaly associated with the front is affected to a much lesser degree as it moves across the mountain.

- When the blocking is strong, the surface front remains trapped on the windward slope and the frontal propagation is discontinuous across the ridge.

During frontal passage over a mountain, several interesting phenomena can occur, such as the development of a lee trough or of a secondary trough along the lee slope, the separation of the upper-level and lower-level frontal waves, and the coupling of the upper-level frontal wave with the secondary trough in the lee of the mountain.

The passage of three-dimensional fronts over mesoscale mountains is more complicated than that of a two-dimensional front. Figure 5.39a shows a schematic of a cold front passing over the Alps, in which the northeastern portion of the front accelerated eastward, while the southwestern portion of the front was retarded at least initially.

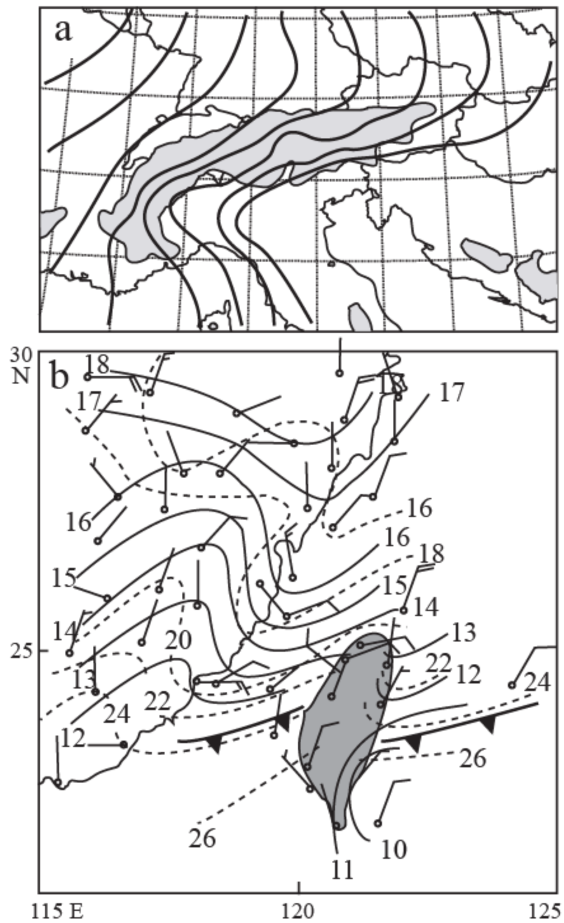

Fig. 5.39: Examples of three-dimensional frontal passage over mesoscale mountains: (a) Schematic of 3-h surface isochrones of a cold front passing over the Alps (shaded). Note that as the front passes over the mountain range it is turned anticyclonically. (Adapted after Steinacker 1981); and (b) Surface analysis of a Mei-Yu front passing over the CMR (shaded) of Taiwan at 1800 UTC 13 May 1987. Solid and dashed curves denote the sea level pressures (mb; variation from 1000 mb) and isotherms (ºC), respectively. (Adapted after Chen and Hui 1992)

In other words, the front passed over the mountain anticyclonically. This type of deformation of the surface fronts has also been observed for Mei-Yu fronts passing over the Central Mountain Range (CMR) of Taiwan (Fig. 5.39b).

The Mei-Yu (Baiu or Changma) front is a nearly stationary, east–west-oriented weak baroclinic zone in the lower troposphere occurring from mid- to late spring through early to midsummer, and typically stretching from eastern China to Taiwan, Japan or Korea.

When an east-west oriented Mei-Yu front passes over the CMR (which is oriented from north-northeast to south-southwest) from the north, the Mei-Yu front at the surface splits into eastern and western flanks, with an eastern flank that accelerates and a western flank that is retarded. In addition to the mountain-induced anticyclonic circulation, the retardation of the left flank of the Mei-Yu front is enhanced by the southwesterly monsoon.

Recent nonlinear numerical experiments are able to: (a) reproduce the orographically induced high pressure, front splitting, and anticyclonic circulation during the passage of Mei-Yu front over the CMR; (b) show the important contribution of the nonlinear advection, Coriolis force, pressure gradient force, and friction to the local rate of change of horizontam momentum; and (c) indicate that the front bears no dynamical resemblance to either an orographically trapped density current or a Kelvin wave (Sun and Chern 2006).

The dynamics of orographic distortion of fronts can be understood by taking a passive scalar approach, in which an initially straight front comes under the influence of the anticyclonic circulation induced by the mountain, and experiences an anticyclonic turning. This is analogous to the passage of a cyclone over a high mountain, as discussed in Section 5.5.

As mentioned earlier, as soon as the front impinges on the mountain, an anticyclonic circulation is produced in the cold postfrontal air and the front advances faster on the left (facing the frontal movement) side of the mountain and is retarded on the right side.

Consequently, weakening of the front (*frontolysis*) is dominant on the left side and lee side facing the direction of frontal propagation, while strengthening of the front (*frontogenesis*) occurs on the right side.

It is found that the degree of blocking on a front along the upwind slope is determined by the Froude number ($F = U/Nh$) and Rossby number (R_o

$= U/fa$, where a is the half-width or horizontal scale of the mountain) (e.g., Davies 1984; Gross 1994).

12.2 Coastally trapped disturbances

A *Coastally trapped disturbance* (*CTD*) is a mesoscale disturbance associated with a high pressure surge along a coastal mountain range, confined laterally against the mountain barrier and vertically by stratification, and propagating along the mountain such that the mountain is on the right in the Northern Hemisphere.

On the U.S. west coast, a CTD forms in spring and summer and propagates northward, and appears as a transition from the prevailing northerly flow. The CTD has a length scale of about *1000 km* along the shore, *100 km* across the shore, and *0.5 km* in the vertical. It normally lasts for 2 to 3 days.

Figure 5.40 illustrates the formation and evolution of CTD along the U.S. west coast.

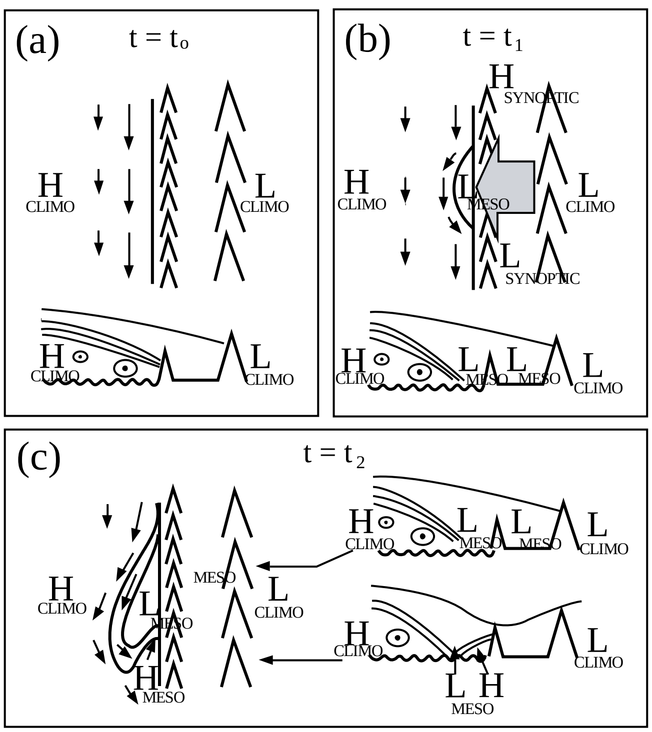

Fig. 5.40: Schematic of an idealized coastally trapped disturbance (CTD) formation and evolution. See text for detailed description. In the figure, "CLIMO" stands for climatological basic state, while "SYNOPTIC" and "MESO" stand for synoptic and mesoscale states, respectively. (Adapted after Skamarock et al. 1999)

Formation and Propagation Processes of a CTD:

- Before the CTD event, the climatological basic state consists of high pressure in the north-central Pacific and a sloping marine boundary layer accompanied by a northerly jet that has its maximum amplitude at the coast, where the boundary layer intersects the coastal mountains (Fig. 5.40a).

- At the early stage of the CTD event (Fig. 5.40b), there is eastward movement of the northcentral Pacific high onshore to the Pacific Northwest, possibly accompanied by westward movement of the low in southwest U.S.

  This synoptic evolution produces offshore flow and a mesoscale coastal low which evacuates the marine layer.

- As the low-level flow comes into geostrophic balance with the mesoscale low, the westerly flow to the south of the low encounters the coastal mountains and elevates the marine boundary layer. This produces a new mesoscale high to the south of the mesoscale low and pushes the marine boundary layer northward as a trapped disturbance (Fig. 5.40c), producing coastal southerlies as the coastal low becomes elongated and is displaced offshore.

During its initial or formation stage, as described above, a CTD behaves like a Kelvin wave (Skamarock et al. 1999). A *Kelvin wave* is a low-frequency inertia-gravity wave trapped in a lateral boundary, such as a mountain barrier in the atmosphere or a continental shelf in the ocean, which propagates counterclockwise in the Northern Hemisphere around a basin.

In the later stage, the CTD is transformed into a density current due to diurnal radiative forcing and differential heating across the coast (Reason et al. 2001). The transition to density current in the later stage might also be affected by other factors, such as nonlinearity and mixing.

A CTD is also known as an *orographic jet* or *ducted coastal ridging* in meteorology and is called *coastally trapped density current* or *coastal jet* in oceanography.

CTDs have also been observed in southeastern Australia, southern Africa, and the southwest coast of South America (Holland and Leslie 1986). Ahead of the coastal ridging in Australia, there often exists a surge of cold air with squally winds. This phenomenon is called the *southerly buster*.

12.3 Cold-air damming

When a cold anticyclone is located to the north of an approximately north-south oriented mountain range in winter, a pool of cold air may become entrenched along the eastern slope and form a cold dome capped by a sloping inversion underneath warm easterly or southeasterly flow. This phenomenon has been observed over the Appalachian Mountains and the Front Range of the Rocky Mountains.

Figure 5.41a shows the surface geopotential temperature field at 1200 UTC 22 March 1985 at the mature phase of a cold-air damming event that occurred to the east of the U.S. Appalachian Mountains.

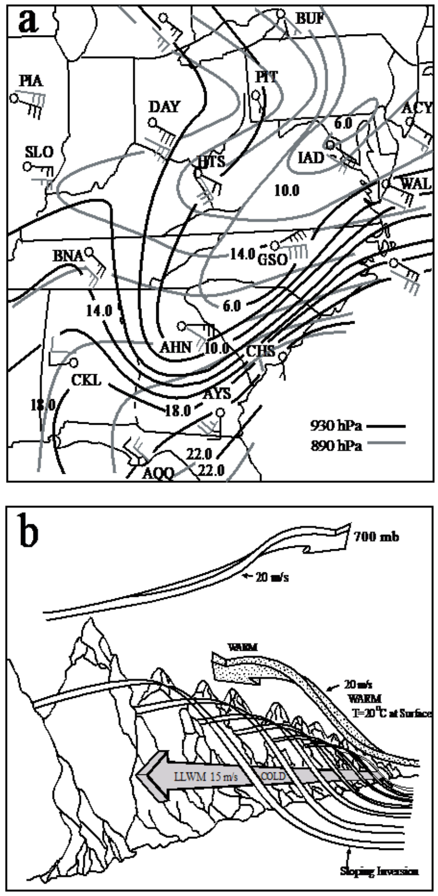

Fig. 5.41: (a) The 930 hPa height (in m, dark solid) and 890 hPa potential temperature (°C, grey solid) fields valid at 1200 UTC 22 March 1985. Winds are in ms^{-1} with one full barb and one pennant representing 5 ms^{-1} and 25 ms^{-1}, respectively. (b) Conceptual model of a mature cold-air damming event. LLWM stands for low-level wind maximum. (Adapted after Bell and Bosart 1988)

- In the initiation phase, a surface low pressure system moved from the Great Lakes northeastward and the trailing anticyclone moved southeastward to central New York.

- The cold, dry air was then being advected by the northeasterly flow associated with the anticyclone. The air parcels ascending the mountain slopes experienced adiabatic cooling and formed a *cold dome*, while those over the ocean were subjected to differential heating when they crossed the Gulf Stream toward the land.

- At the mature phase, the surface anticyclone remained relatively stationary in New York and the cold air was advected southward along the mountain slopes within the cold dome.

Figure 5.41b shows a conceptual model of the cold-air damming occurred to the east of the Appalachian Mountains at the mature phase.

The low-level wind (LLWM in the figure) moved southward at a speed of about 15 ms^{-1} within the cold dome. The cold dome was capped by an inversion and an easterly or southeasterly flow associated with strong warm advection into the warm air existed above the dome. Moving further aloft to 700 mb, the wind flows from south or southwest associated with the advancing short-wave trough west of the Appalachians.

As described above, the cold-air damming process can be divided into the initiation and mature phases.

- During the initiation phase, the low-level easterly flow adjusts to the mountain-induced high pressure and develops a northerly barrier jet along the eastern slope of the mountain range (Fig. 5.41b), similar to that shown in Fig. 5.36a except for the basic flow direction.

  Because the upper-level flow is from the southwest, it provides a cap for the low-level flow to climb over to the west side of the mountain. In the meantime, a cold dome develops over the eastern slope due to the cold air supplied by the northerly barrier jet, adiabatic cooling is associated with the upslope flow, and/or the evaporative cooling.

- In the mature stage, the frictional force plays an essential role in establishing the steady-state flow with the cold dome (Xu et al. 1996). It was found that the cold dome shrinks as the Froude number (U/Nh) increases or, to a minor degree, as the *Ekman number* ($\nu/(fh^2)$, where ν is the coefficient of eddy viscosity) decreases and/or the upstream inflow veers from northeasterly to southeasterly. The northerly barrier jet speed increases as the Ekman number decreases and/or the upstream inflow turns from southeasterly to northeasterly or, to a lesser degree, as the Froude number decreases.

12.4 Gap flow

When a low-level wind passes through a gap in a mountain barrier or a channel between two mountain ranges, it can develop into a strong wind due to the acceleration associated with the pressure gradient force across the barrier or along the channel.

Gap flows are found in many different places in the world, such as the Rhine Valley of the Alps, Senj of the Dinaric Alps, Independence, California, in the Sierra Nevada, and Boulder, Colorado, on the lee side of the Rockies (Mayr 2005).

Gap flows occurring in the atmosphere are also known as mountain-gap wind, jet-effect wind or canyon wind. The significant pressure gradient is often established by (a) the geostrophically balanced pressure gradient associated with the synoptic-scale flow and/or (b) the low-level temperature differences in the air masses on each side of the mountains.

Based on Froude number ($F = U/Nh$), three gap-flow regimes can be identified:

(1) Linear regime (large *F*): with insignificant enhancement of the gap flow;
(2) Mountain wave regime (mid-range *F*): with large increases in the mass flux and wind speed within the exit region due to downward transport of mountain wave momentum above the lee

slopes, and where the highest wind occurs near the exit region of the gap; and

(3) Upstream-blocking regime (small *F*): where the largest increase in the along-gap mass flux occurs in the entrance region due to lateral convergence (Fig. 5.42; Gaberšek and Durran 2004).

Gap flows are also influenced by frictional effects which imply that: (1) the flow is much slower, (2) the flow accelerates through the gap and upper part of the mountain slope, (3) the gap jet extends far downstream, (4) the slope flow separates, but not the gap flow; and (5) the highest winds occur along the gap (Zängl 2002).

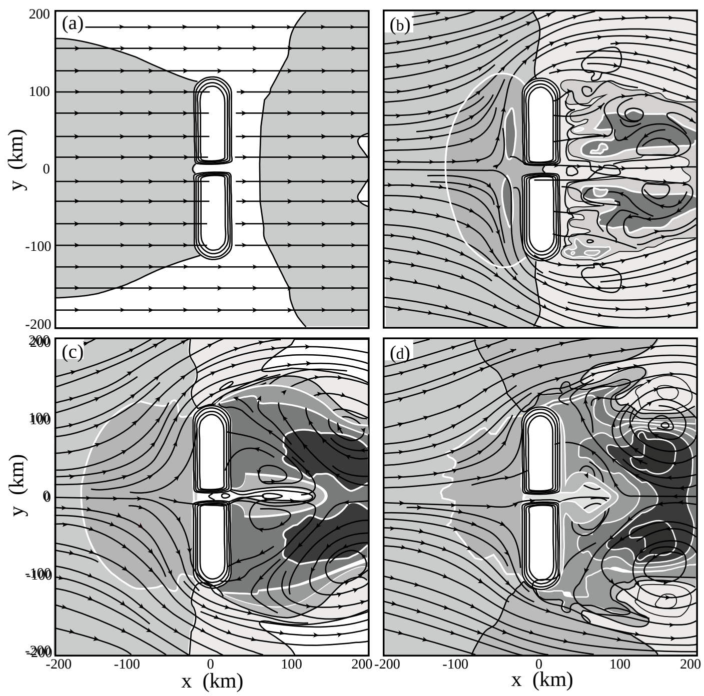

Fig. 5.42: Horizontal streamlines and normalized perturbation velocity $(u-U)/U$ at z = 300 m and Ut/a = 40, for flow over a ridge with a gap when the Froude number (U/Nh) equals (a) 4.0, (b) 0.72, (c) 0.36, and (d) 0.2. The contour interval is 0.5; dark (light) shading corresponds to negative (positive) values. Terrain contours are every 300 m. (From Gaberšek and Durran 2004)

12.5 Dynamics of mountain-plains solenoidal circulations

- The dynamics of *mountain-plains solenoidal (MPS) circulations* is a little explored area of orographically influenced flow and weather phenomena. This is mainly due to the complicated interactions between orographic and thermal forcings.

- Taking into consideration sensible heating or cooling over elevated terrain results in a considerably more complex flow than has been considered until now. The classical view of orographically and thermally forced winds in mountains includes the slope and mountain-valley winds.

  - During the day, the mountain serves as an elevated heat source due to the sensible heat released by the mountain surface.

    In a quiescent atmosphere, this can induce mountain *upslope flow or upslope wind*, which in turn may initiate cumuli or thunderstorms over the mountain peak and produce orographic precipitation.

  - At night, the opposite occurs: surface cooling produces downslope *drainage flow*.

- Based on observations, four stages in the development of a thermally forced circulation generated by solar heating in a mountain valley have been identified (e.g., Banta 1990):

I. Before sunrise, the nocturnal inversion layer contains drainage flow, which generally blows in a different direction from the winds above the inversion. Just prior to sunrise, this very stable layer remains adjacent to the surface;
II. After sunrise, surface sensible heating erodes the inversion layer and produces a shallow *convective boundary layer* (*CBL*) below the inversion layer and the upslope flow;
III. The shallow CBL or upslope layer deepens as the surface heating continues; and
IV. After the nocturnal inversion layer disappears during the afternoon, a deep, well-mixed CBL is created.

Linear theories described in Sections 6.1 and 6.2 have been applied to study the combined effects of orographic and thermal forcing for mesoscale mountain flow (e.g., Raymond 1972; Smith and Lin 1982). Numerical modeling studies of the combined orographical and thermal forcing have been explored as early as the 1960's (e.g., Orville 1964, 1968). More sophisticated numerical models with a variety of initial conditions have been adopted in the more recent studies of mountain-plains solenoidal circulations. The results given by these models have been verified by conventional observations as well as field experiments (e.g. Tripoli and Cotton 1989; Wolyn and McKee 1994).

- Figure 6.26 shows a conceptual model for the daytime evolution of the MPS circulation.

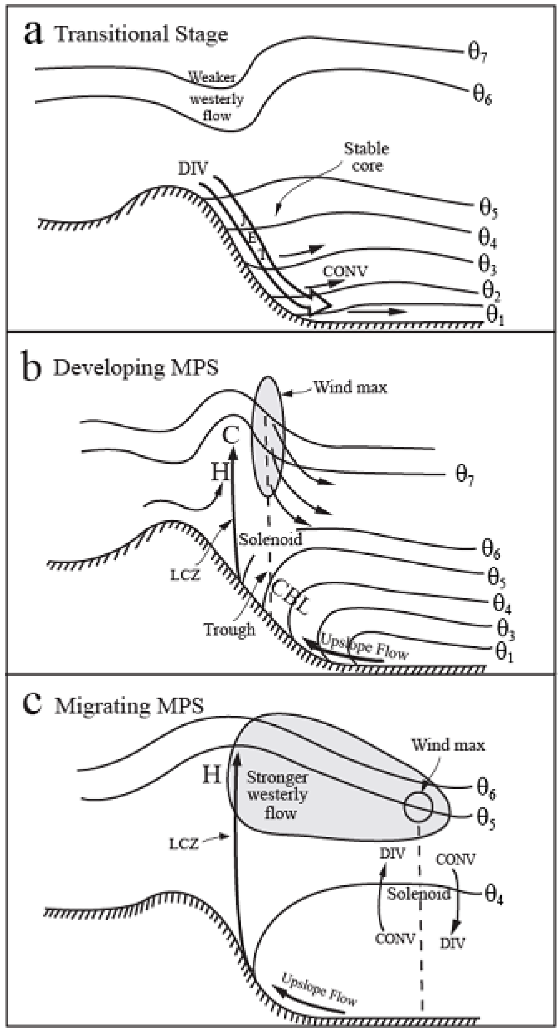

Fig. 6.26: Conceptual model of the daytime evolution of the mountain-plains solenoidal (MPS) circulation east of a mesoscale mountain under conditions of clear skies, steady-state synoptic-scale situation, and light basic westerly wind (e.g., 5 ms^{-1}). Three primary stages may be identified: (a) transition stage, (b) developing MPS stage, and (c) migrating MPS stage. Symbols DIV, CONV, JET, C, H, CBL, and LCZ denote divergence, convergence, katabatic jetlike flow, cold core, higher pressure, convective boundaru layer, and leeside convergence zone, respectively. Regions of wind maximum are shaded. Solid lines are the isentropes. (Adpated after Wolyn and McKee 1994)

The circulation primarily includes:

- Transitional stage
- Developing mountain-plain solenoidal (MPS) stage

- Migrating MPS stage.

The transitional stage occurs when the sun rises. The most pronounced feature of the transition stage is the katabatic jetlike flow down the east side of the mountain (Fig. 6.26a).

The slowing of the nocturnal jet on the eastern plains produces a convergence that lifts the cold air, thus creating a *stable core* that is shallower farther east of the barrier. This nocturnal katabatic flow weakens as it is affected by the surface heating, and is replaced by a mesoscale solenoidal circulation 3-4 h after sunrise (Fig. 6.26b).

A shallow convective boundary layer (CBL) is produced below the inversion layer and an upslope flow is produced by the horizontal pressure gradient force toward the slope in response to the buoyancy associated with the surface sensible heating. The main upward motion of the solenoidal circulation occurs in a narrow zone over the eastern slope of the mountain, and is called the *leeside convergence zone* (LCZ). The LCZ lifts the air into the ambient air above, creating the cold core (denoted by "C" in Fig. 6.26b).

A strong sinking motion occurs to the east of the cold core, creating a pressure trough in which the center of the solenoid is located. The horizontal pressure gradient associated with the cold core and the trough to the east produces a horizontal wind speed maximum. A broad region of sinking motion is located to the east of the solenoid center.

At the later time of this stage, the sinking and horizontal warm-air advection immediately east of the solenoid center is able to warm the air enough to create a negative pressure gradient in the stable core above the CBL.

The final stage of the mountain-plains solenoidal circulation is characterized by the eastward migration (Fig. 6.26c). Convergence (divergence) near the height of the wind maximum region and divergence (convergence) near the surface tend to produce sinking motion ahead (behind) the horizontal wind maximum located beneath the leading edge of the cold core.

The solenoid center is located in a pressure trough beneath the eastward-moving leading edge of the cold core, while the LCZ remains anchored over the lee slopes. Only the migrating MPS may be defined as a disturbance, and as thus can significantly affect the atmosphere on the plains located east of the system during the daytime circulations. The CBL grows explosively and the depth of the upslope flow increases when the solenoid passes a location.

The MPS has been shown to be responsible for producing a strong updraft, which in turn generated the dominant wave of the second episode of gravity waves observed on 11-12 July 1981 during the Cooperative Convective Precipitation Experiment (Koch et al. 2001). A gravity wave was generated as the updraft impinged upon a stratified shear layer above the deep, well-mixed boundary layer developed by strong sensible heating over the Absaroka Mountains. Explosive convection developed directly over the remnant gravity wave as an eastward-propagating density current, produced by a rainband generated within the MPS leeside convergence zone, merged with a westward-propagating density current in eastern Montana. The complicated interactions of differing sensible heat contributions from complex terrain, gravity waves, and convection indicate the need for increasingly detailed observations and theories to verify existing MPS hypotheses and gravity wave generation.