Mesoscale effects and PBL

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Many mesoscale phenomena and their PBL are closely intertwined. The complex PBL provides the horizontal differences, which drive the mesoscale circulations, the local winds and temperatures of which then control the PBL structure. We describe here briefly two classes of surface-bound mesoscale circulations: thermally driven and mechanically driven. Excluded are e.g. frontal and deep convective phenomena, they are also closely related to their PBL's. Stull's textbook (1988) provides a good introductory text, from which most of the figures below are taken.

1. 1 Thermally driven circulations; heat islands

Whenever the surface is relatively warm (e.g. an open sea in winter, or land in summer sunshine), convection and LW radiation quickly carry the heat to the air. The PBL warms up, it expands, and creates (through hydrostatics) a local high pressure at the top of the BL. Pressure gradient force then accelerates air out of this warm upper high, being replaced by rising warm air from below (by continuity). Hence a 'thermal low' is created over the warm surface. Cooler air from the sides then flows into the thermal low. In larger scales Coriolis force creates cyclonic inflow near the surface, and anticyclonic outflow near the top of the convective BL (CBL) over the heated area.

This mechanism operates in many scales. Summer monsoons over hot subtropical continents are a large-scale example. Sea breezes blow over sunny coastal zones. The urban heat island is yet another example, as are ubiquitous slope winds over slopes and hilly terrain. An open sea gulf or lake may cause heavy snowfall on the downwind coast due partly to the heat island -induced rising motion. All-sized islands in the ocean are often marked by cumulus clouds induced by the rising motions.

In calm weather the heat island creates thermal circulation cell(s), where air from all cool sides converges into the thermal low over the warm zone; the warmed air rises, diverges out of the thermal high at the top of the CBL, cools (by LW radiation) and sinks.

If there is prevailing wind, the upwind cell is advected over the warm zone and gets phase-locked to it: A warm plume of rising air develops downstream and sinking motion in front with mass-conserving wind anomalies: strong wind just above and weak wind aloft. The stronger wind enhances heat transfer to the air. There is a lot of turbulence and entrainment downstream near the top of the CBL. Thermal standing gravity waves are excited as well. For a review, see Savijärvi and Matthews (2004 JAS).

During clear nights, real islands are typically slightly cooler than the surrounding ocean. Then, a nocturnal reverse 'anti-heat-island' circulation may form but it remains weak and shallow because of small temperature differences and stable stratification.

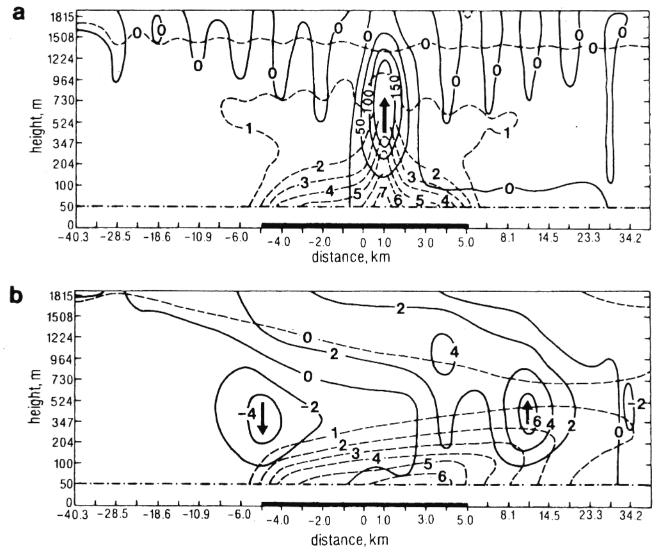

Fig. 23. Development of a local circulation over a heated island (*a*) with $\bar{u}$ = 1 m s^{-1} and (*b*) with $\bar{u}$ = 5 m s^{-1}. In both cases ΔT = 10°C, and the simulation is for 5 hours after onset of island heating. Solid lines indicate vertical velocity perturbations in centimeters per second, and broken lines temperature disturbance in degrees Celsius [after *Tanouye*, 1966].

(note the rising and sinking motion and warm plume in b)

1.2 Urban heat island

gets its heat through industrial and domestic heating, traffic, solar heat stored in buildings and released at night, many black, dry, black surfaces (roads, parking plots) etc. A city can be 1-2 K warmer than the surrounding countryside during the day, and up to 10-12 K warmer overnight.

In calm weather an urban heat island circulation (UHIC) is induced, which is strongest in the afternoon when the surface winds toward the city are however only 1-2 m/s. During the night strong stability and calm weather damp turbulence everywhere. Hence even a large temperature difference cannot drive UHIC effectively during the night, and the winds die away. A surface inversion may therefore develop, with pollution.

In windy weather UHIC forms an 'urban plume' on the downwind side. The associated rising motions within the warm, dry and polluted plume may here prevail day and night, because wind now maintains turbulence and transfers the surface temperature differences (and materia) to the air. Hence for instance thunderstorms, rain and snowfall may be locally enhanced on the downwind side of a city. The wind-enhanced convection over the urban heat island may also increase surface winds and reduce their cross-isobar angles, so that winds often turn anti-cyclonically as their enter the city area or other heat islands.

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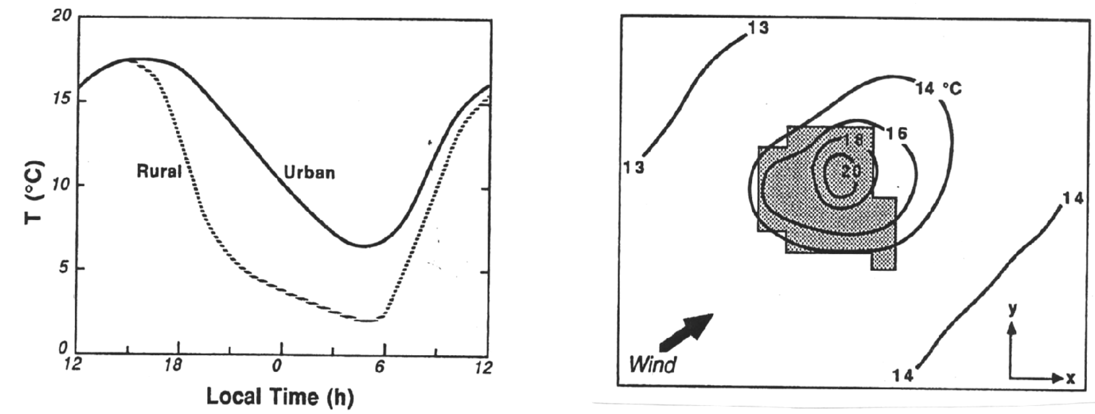

The typical 2m temperature curves in the city centre and countryside (left), and a map of the typical evening 2m temperature distribution in a city area (right).

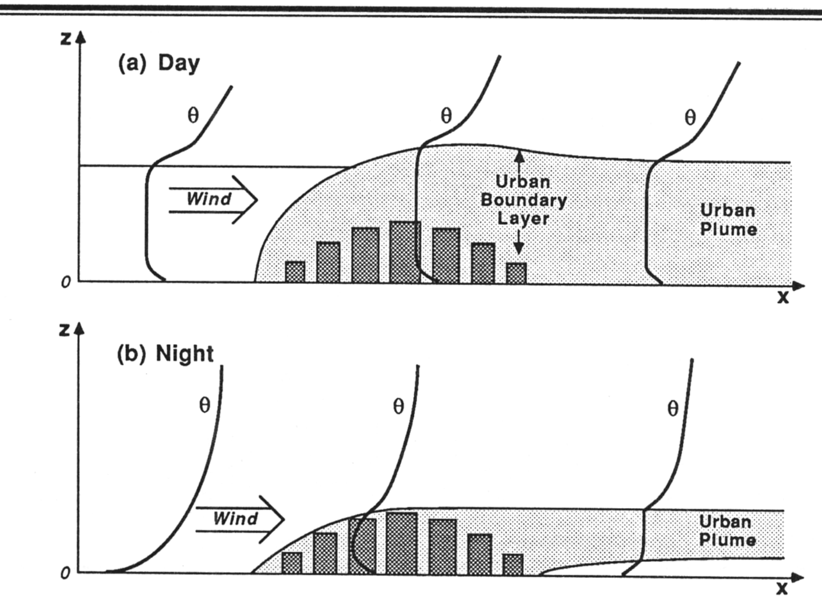

Fig. 14.22 Sketch of the urban boundary layer and urban plume for a windy (a) day, and (b) night. (After Oke, 1982).

Notes:

- UHIC prevails within other flows. Therefore a warm coastal city may e.g. locally enhance sea breezes and weaken land breezes (Savijärvi, 1985 Geophysica).

- In valley bottom cities (e.g. Lanzhou, Mexico City) there is a danger ('smog trap'): During a calm day, the upslope wind cells and the UHIC cells are opposite and kill each other, so there is no local wind. During the night, downslope winds together with UHIC keep advecting pollutants to the city centre, where they are trapped in the stable nighttime BL (Savijärvi and Jin, 2001 BLM).

1.3 Sea and land breeze

During a clear, calm day a heat-island and thermal low may develop over sunny land. This drives sea breeze circulation over coasts as the BL air above cold water remains cool. The afternoon surface pressure difference across a narrow coastal zone is about 1 mb. The SB circulation cell is 50-60 km long and 1-2 km high. It proceeds inland at about 3 m/s in calm air. The cool, moist surface winds are 3-7 m/s from the sea and the advancing 'sea breeze front' of rising motions is often marked by a wall of cu or cb clouds. Winds are slowly turned by Coriolis force and may change to a weaker land breeze overnight, when the land is typically slightly cooler than the sea.

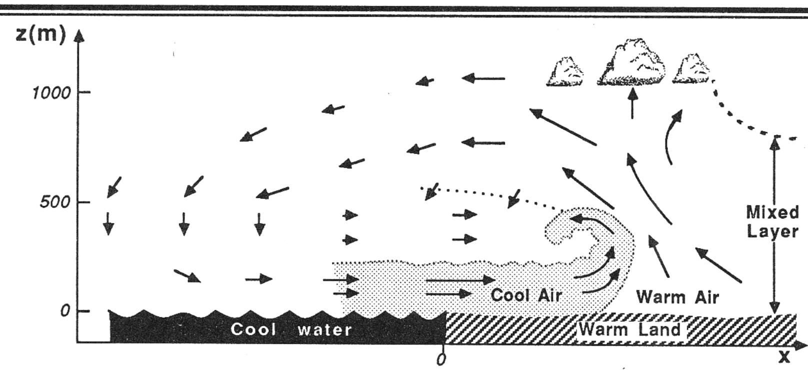

Fig. 14.7 Idealized sea or lake-breeze circulation. (After Lyons (1975) and Ogawa, et al., 1986).

The temperature difference between land and sea air is crucial, 8 K or more is needed for a good onset of SB. In cloudy air SB does not exist. Prevailing wind from the sea advects the SB cell quickly inland where it disappears, while weak wind from land keeps the cell over the coastline, where it gains strength throughout the day. If wind from land is > 7 m/s, the SB cell may not form or may be advected (pushed) out to the sea.

On the Gulf of Finland, moderate southerly winds may push the strong Estonian SB cell over to Finland, causing easterly coastal winds (Savijärvi and Alestalo, 1988 BPA). During moderate southeasterly winds, sea breezes on the south coast of Finland may combine with inertial oscillations (the well-mixed continental air entering the smooth Gulf, accelerating and advecting over the sea). This may lead to supergeostrophic surface easterlies along the coast. Both cases are associated with an easterly low level jet, due to the strongly varying BL structure between land and sea (Savijärvi et al., 2005 QJ).

Notes:

- On steep coasts the sea breeze may jump straight uphill from the sea, creating calms or even countercurrents over the coastline (Neumann and Savijärvi, 1986 BPA).

- Sea breeze is a systematic mesoscale mixing mechanism at the coasts, which should perhaps be parameterized in climate models. It is well resolved only with horizontal gridlengths of about 5 km or less (Savijärvi, 1995 BPA).

- In sunny tropical coasts and lakes sea breezes are common. Slope-assisted land breezes may also exist and may trigger nighttime thunderstorms onto lakes, e.g. over Lake Victoria and Lake Tanganyika in Africa (Savijärvi, 1997 QJ).

- An island at least 40 km wide develops sea breezes at full strength. In a larger island or peninsula the advancing sea breeze cells from the opposite coasts may converge in the middle, creating afternoon thunderstorm lines (Savijärvi and Matthews, 2004 JAS).

- Over the boundary between wet, evaporating cool land and dry hot land, a sea-breeze –like 'moist breeze' may develop. A snow boundary may create a 'snow breeze' - the melting snow remains at 0 C while the bare land may be much hotter in sunshine.

1.4 Slope winds

Over a sunny slope (even 1:1000) the convectively heated BL air is warmer than air away from the slope at the same height. This temperature difference drives a warm uphill wind, which is weak but exists in the deep CBL. During a clear night, the ground cools effectively, the air cools, too (by turbulence and LW radiation), and the dense air flows downslope. These nighttime <u>katabatic</u> winds are common in all hilly landscapes. They are shallow but stronger than the daytime uphill winds. The landscape guides the heavy, cool and very stable flow strongly. The flow may continue downstream even after the slope ends (by its own momentum, and because friction is small in the stable slow air mass).

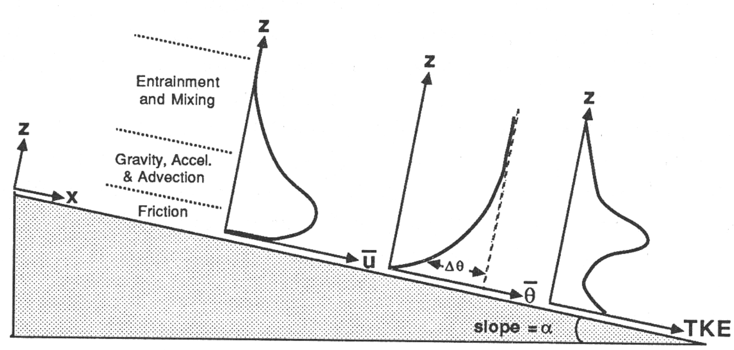

Fig. 12.23 Downslope drainage winds in a SBL.

In longish valleys there may also develop a similar but longitudinal warm 'up-valley wind' during daytime, and cold 'mountain wind' during nighttime. In a hilly area the local katabatic flows bring the heavy air into the valley bottoms, which are thereby filled by stagnant cold air. On fjell slopes, for instance, there may well be windy and 10-20 K warmer than at the calm cold valley bottoms during these surface inversion cases.

Over the long, steep, icy and cold slopes of the Antarctic and Greenland katabatic flows may grow strong and very steady. They are turned by the strong Coriolis force. Advection and conservation of mass may accelerate all drainage flows. This can lead into 'shooting flows' and bora in mountaineous terrain (see next section).

Notes:

- The upslope wind is weak in a narrow valley as the whole air mass is quickly warmed up by recirculation. The downslope wind is less sensitive to the valley width but being stronger, it is sensitive to turning by Coriolis force and hence latitude (Savijärvi and Jin, 2001 BLM).

- On a sloping coast upslope winds may assist sea breezes during daytime, and katabatic winds land breezes during the night (Mahrer and Pielke 1977 MWR, Savijärvi and Matthews, 2004 JAS).

- The U.S. Midwest is sloping steadily from Missisippi up to the Rocky Mountains. This creates easterly upslope winds during daytime and westerly katabatic winds during the night. They may increase the southwesterly nocturnal boundary layer jet (NBJ) by about 25% in the typical case (Savijärvi, 1991 MWR).

- An east-facing slope, southwesterly basic flow and strong diurnal variation is characteristic also in the Viking Lander 1 site in Mars, so slope winds and a strong NBJ are expected there as well (Savijärvi and Siili, 1993 JAS)

2.1 Mechanically driven mesoscale phenomena, stable BL

Flow over roughness changes can cause e.g. small-scale coastal convergence (with mechanically-induced rising motions, Savijärvi, 2004 Tellus), and multiple internal boundary layers in complex terrain. The rougher elements tend to dominate turbulence in a mosaic of rough and smooth surfaces (Vihma and Savijärvi, 1991 QJ).

We concentrate however here on the effects of topographical variations only, without surface temperature or roughness differences. If a stable flow with speed U and Brunt-Väisälä frequency N climbs over a hill with width L, the natural horizontal wavelength of the induced gravity oscillation is $2\pi U/N$ while the forcing wavelength is 2L. Their relation $U/(NL)$, called the Froude number, governs the flow response.

If Fr << 1 the air is very stable and heavy: it turns around or makes a blocking wedge in front of a hill rather than climbs over it. Cold air pools fill the valleys and the flow is strongly guided around all obstacles in the 0-10 km scale. (This is the typical case in Lapland and at the Finnish coasts up to June: the fjells and the islands guide the very stable flow strongly while stability is maintained by the cool surface.)

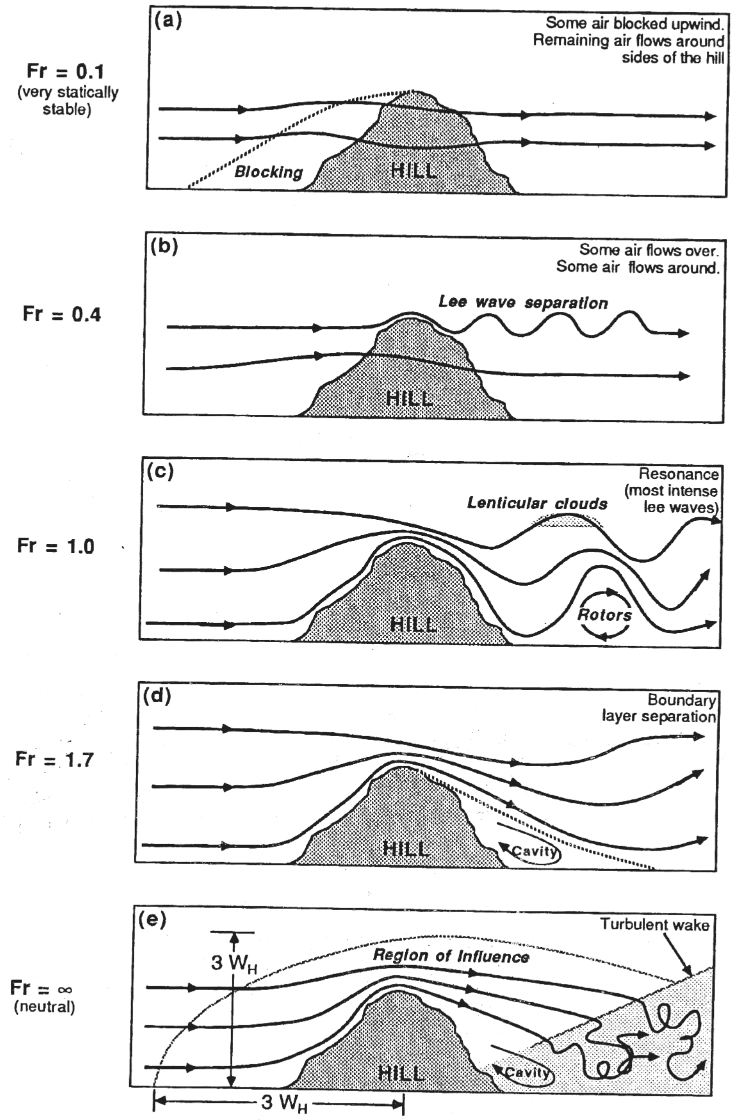

Fig. 14.14 Idealized flow over an isolated hill. The Froude number (Fr) compares the natural wavelength of the air to the width of the hill (W_H).

If Fr is about 1, there is resonance, and a strong standing gravity wave ('mountain wave') pattern is triggered, with upstream-tilting phases and strong, gusty winds observed behind the hill ('downstream windstorms'). At Fr 0.1…0.9 the waves may exist near the mountain top but not near the surface (lee wave separation).

At Fr > 1.5, the wind behind the hill is slow and a cavity flow separation may form while wind at the hilltop is now quite strong.

For small hills with L less than about 5 km the flow just adapts locally, and the effect of the hill decays rapidly upward.

2.2 WKB solutions, vertically variable but stable BL

The mountain waves of a linear model are governed vertically by an oscillation equation (1) $w_{zz} + m^2 w = 0$, where $m^2 = N^2/U^2 - k^2$. The horizontal wave number k = 2π/L is fixed by the mountain width L and m is the vertical wave number. The wavy ($m^2 > 0$) solution is w = A exp(imz). This assumes constant N, U, A and m. If we however redefine $w = A(z)\exp(i\int_0^z m(z)dz)$, and substitute it to (1), $\frac{d^2A}{dz^2} + 2im\frac{dA}{dz} + iA\frac{dm}{dz} = 0$ is the result. If A(z) is now varying only slowly, the first term can be neglected as small, and one gets $d(A^2m)/dz = 0$ (the so-called WKB approximation). For instance, if the stability N increases downward (U being constant), m also increases downward but the vertical wavelength $L_z = 2\pi/m$ increases upward. The wave amplitude A(z) must now increase upward (to conserve vertically the product A^2m). Similar results are obtained for upward-increasing U (z), which is the usual case in a baroclinic flow.

The WKB approximation can thus be used to chart the effect of vertically varying BL on standing mountain waves and on other linear model results.

2.3 Well-mixed flow over an obstacle

If a well-mixed CBL flow with a typical capping inversion overpasses a low hill, the flow accelerates at the hilltop (to conserve mass), and the inversion falls down locally by a so-called Bernoulli effect. (it also sucks the shower curtain, the inversion, toward you, the obstacle, when you take a shower). Strong turbulent wind at the hilltop is the result. This is the usual sunny day case e.g. in Lapland.

If the hill/slope is relatively high and wide, the local Froude number may turn supercritical. The downslope flow then accelerates to great speeds and makes a 'hydraulic jump'. This results in very gusty leeside windstorms.

If the hill is higher than the capping inversion, the CBL flow goes around the obstacle and makes Karman vortices behind. These are often seen in satellite pictures in the form of cu/sc cloud vortex streets behind high islands, such as Jan Mayen.

2.4 Föhn and Bora

If a stable wind blows over a large mountain, 1) upstream blocking may cause air to flow from dry midtroposphere down the mountain, or 2) forced uphill rain dries the air mass. The downflowing air warms up adiabatically, and a warm, dry föhn is the result.

If the air is very cold, stable and 'heavy' initially (e.g. a northerly flow over the Alps), it rushes down the mountain gaps and valleys at great speed and stays cold and dry. This gusty, dry wind is known as bora (in Kroatia) and mistral (in Rhone valley), and it is very unpleasant. Both föhn and especially bora may further trigger standing mountain waves and windstorms by the local hills on their way. These may further add to the general wind speed and gustiness.

Summary:

In nature, the quasi-horizontal topographic, roughness and surface temperature differences are all acting together. The result is a rich variety of responsing mesoscale flow patterns, which are closely tied to their BL structures in a two-way interaction, and may themselves be embedded on a larger-scale basic flow, interacting with it.

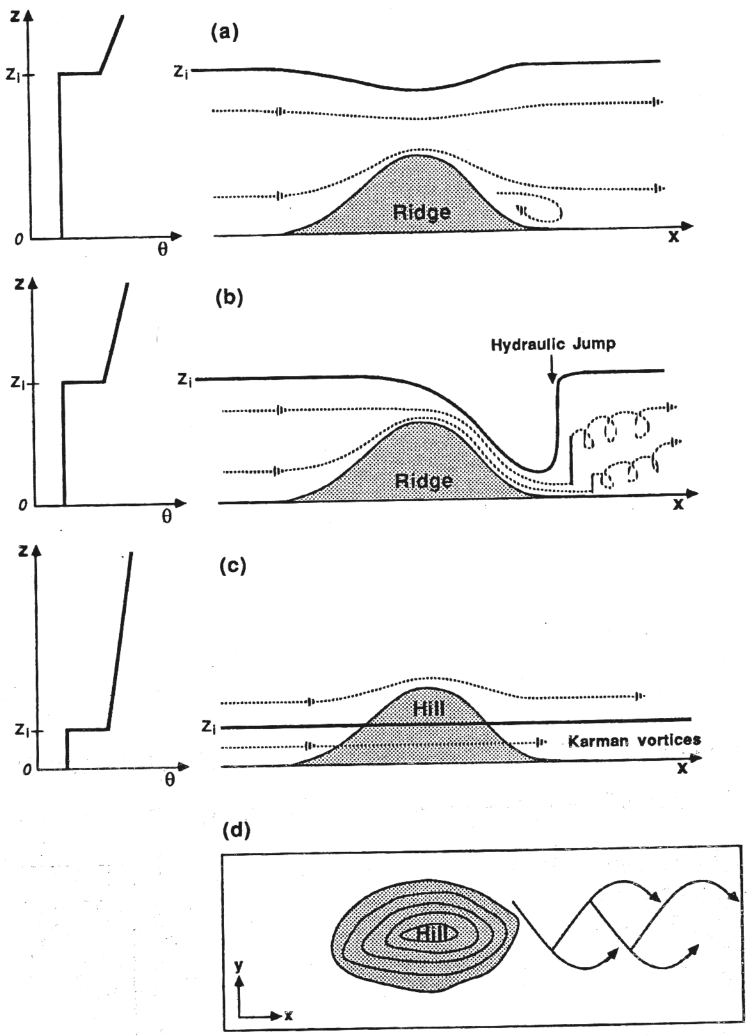

Fig. 14.19 (a) Bernoulli effect as flow accelerates over a 2-D ridge; (b) hydraulic jump downwind of a 2-D ridge for greater ambient wind speed; (c) flow around the sides of an isolated hill; and (d) Karman vortex street downwind of the hill from (c). (After Hunt, 1980.)