GEOG 321 - Reading Package Lecture 28

THERMALLY GENERATED WINDS

The juxtaposition of contrasting thermal environments results in the development of horizontal pressure gradient forces, which if sufficient to overcome the retarding influence of friction will cause air motion across the boundary between the surfaces.

Land and sea breeze circulation. Land and water surfaces possess contrasting thermal responses because of their different properties and energy balances, and this is the driving force behind the *land* and *sea (lake) breeze* circulation system encountered near ocean or lake shorelines. Compared with most land surfaces a water body exhibits very little diurnal change in surface temperature.

The reasons for the development were discussed in GEOB 200 and ATSC 201, but in summary water is different because it (i) allows transmission of short-wave radiation to considerable depths, (ii) is able to transfer heat by convection and mixing, (iii) converts much of its energy surplus into latent rather than sensible heat, (iv) has a large thermal inertia due to its higher heat capacity. Thus although Q^* may be greater over water (because of its low albedo), the effectiveness of Q_E and ΔQ_S as thermal sinks means that Q_H is small. By day Q_H is small because most of the energy is channelled into storage or latent heat; at night it is small because the long-wave radiative cooling is largely offset from the same water store. The reduced convective heat flux (Q_H) to and from the air means that atmospheric warming and cooling rates ($\partial T/\partial t$) are relatively small over water bodies. In contrast the convective fluxes and rates of temperature change over land are large and show a marked diurnal variation.

These land/water temperature differences and their diurnal reversal (by dayland warmer than water; at nightland cooler than water) produce corresponding land/water air pressure differences. These in turn result in a system of breezes across the shoreline which reverse their direction between day and night (Figure 1).

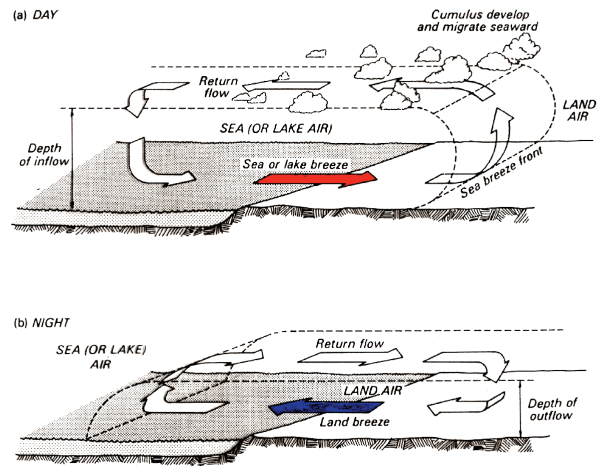

Figure 1: Land and sea (lake) breeze circulations across a shoreline (a) by day and (b) at night, during anticyclonic weather.

In the morning the greater Q_H over the land heats the air column more rapidly and to greater heights than over the water. The consequent expansion of the land column means that the pressure aloft becomes higher than at the same level over the water. This results in a flow at upper levels towards the water, in so doing it must produce greater pressure at the surface over the water and hence a cross-shoreline flow develops from water to land (the sea or lake breeze). The land breeze is initiated in the evening due to the greater cooling and contraction of the air column over the land. Thermal breeze systems of this type are best developed in anticyclonic summer weather because almost cloudless skies and weak synoptic-scale winds permit the maximum differentiation between surface climates. Increased cloud or stronger winds modify or obliterate these local winds. Note that both the land and sea breezes are really the low-level portion of a complete circulation cell.

The daytime sea breeze circulation (Figure 1a) has a greater vertical and horizontal extent, and its wind speeds are higher, than the nocturnal land breeze (Figure 1b). This is because by day the solar forcing function is in operation and instability is greatest. Commonly the sea breeze blows at 2 to 5 $\mathrm{m\,s^{-1}}$, extends inland as far as 30 km, and affects the air flow up to a height of 1 to 2 km. The horizontal extent of the breeze may allow it to be deflected by the Coriolis force so that by late afternoon the inflow ends up being almost parallel with the coast. On the other hand the land breeze is usually about 1 to $2\,\mathrm{m\,s^{-1}}$ in strength and smaller in both horizontal and vertical extent. During the sea breeze the cooler more humid sea or lake air advects across the coast and wedges under the warmer land air. The advancing sea breeze front produces uplift in what is already an unstable atmosphere over the land. The front is therefore commonly associated with the development of sea breeze cumulus clouds which are caught up in the counter flow aloft and are carried seaward where they dissipate because they have been removed from their moisture source, and because over the water the air subsides, and warms adiabatically.

The lowest layer of sea or lake air which is advected across the coast is modified by the 'leading-edge effect'. The stable marine air is warmed over the land producing a more unstable internal boundary layer which grows in depth with distance from the shore (not shown in Figure 1 to avoid clutter). If the coastal water is upwelling and therefore relatively cold it may remain cooler than the land even at night. The lack of temperature reversal may then encourage the sea breeze to continue through the night.

Other thermal circulations. In direct analogy with the sea breeze system a city can generate 'country breezes'. When regional winds are very weak a city is commonly warmer than its surroundings. This induces low-level breezes across the perimeter of the city which converge on the centre. It is found that the vertical temperature structure is at least as important as the urban-rural temperature difference. Vertical instability aids the three-dimensional circulation. By day small horizontal differences are sufficient to drive the system. Unlike the sea breeze system there is no diurnal reversal of flow the city is almost always warmer.

Any major conflagration such as a forest or brush fire, large-scale burning of agricultural or forest debris after harvest etc. can generate convergent winds at its base.

The thermal contrast between the cool sub-canopy interior of a forest by day, and the un-shaded surrounding fields or grassland, can promote cool breezes blowing out from the stand border. A reverse flow should occur at night but it seems that the frictional drag of the forest restricts its effects to a few metres in from the edge. In fact on good radiation nights it is more common to encounter cool breezes moving away from the forest edge. This is air draining intermittently down from the forest canopy because that is the elevated active surface.

TOPOGRAPHICALLY GENERATED WINDS

Valleys, especially those in mountainous regions, produce their own local wind systems as a result of thermal differences. As with the land and sea breeze thermal circulation the local winds of valleys are best developed in anticyclonic weather in summer. Under such conditions, with almost cloudless skies and weak large-scale motion, differential warming or cooling of different facets of the landscape gives rise to horizontal temperature and pressure gradients, which cause winds. The exact nature of these wind systems depends on the orientation and geometry of the valley. The best developed and most symmetric wind system might be anticipated in a deep, straight valley with a north-south axis. In valleys with other orientations or possessing complex geometries (e.g. bends, constrictions, etc.) the flow pattern may be asymmetric or incomplete. For convenience we will consider the case of a simple north-south valley, but even then there must be some asymmetry with time due to the diurnal variation of solar radiation input to west- and east-facing slopes.

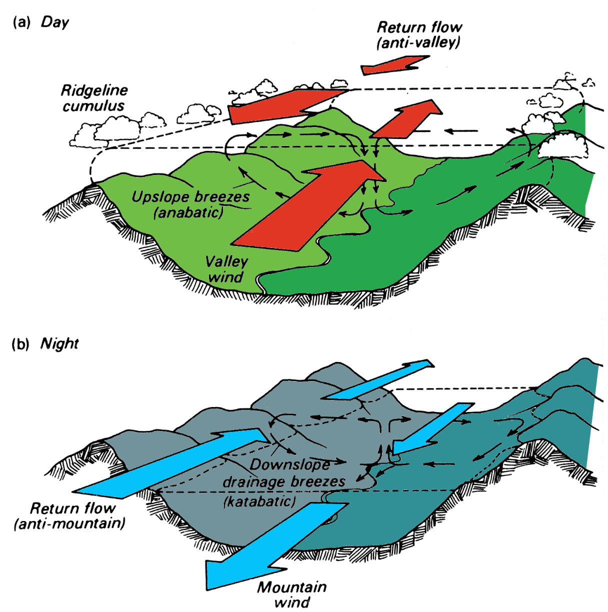

Figure 2: Mountain and valley wind system viewed with the reader looking up-valley. (a) By day slope winds are anabatic, and the valley wind fills the valley and moves upstream with the anti-valley wind coming downstream, (b) At night the slope winds are katabatic and reinforce the mountain wind which flows downstream, with the anti-mountain wind flowing in the opposite direction above.

Anabatic and valley-wind. By day the air above the slopes and floor of the valley will be heated by the underlying surface to a temperature well above that over the centre of the valley. As a result shallow, unstable upslope (or *anabatic*) flow arises, and to maintain continuity a closed circulation develops across the valley involving air sinking in the valley centre (Figure 2). Commonly the uplift along the slopes is at speeds of 2 to 4 $\mathrm{m\,s^{-1}}$ with a maximum at about 20-40 m above the surface. It can lead to the formation of convective anabatic clouds along the valley ridges. In tropical valleys this may lead to greater precipitation along ridges in comparison with the valley floor. The cross-valley circulation also effectively transports sensible heat (Q_H) from the surrounding active surfaces to warm the whole valley atmosphere. Therefore when compared to the atmosphere at the same level over an adjacent plain (or further downstream) the valley air is much warmer, and in a manner analogous to the sea breeze, a plain-to-mountain flow develops. The up-valley flow is termed the *valley wind*, and fills the entire valley. The maximum pressure gradient is near the surface, and hence the maximum wind speed is as close to the ground as the retarding influence of the surface allows. Above the ridges there is a counter flow (the anti-valley wind) which flows down-valley by day (Figures 2a and 3). Again the similarity with the sea breeze circulation cell is evident. Above the *anti-valley wind* there will be yet another wind associated with the large-scale synoptic flow pattern.

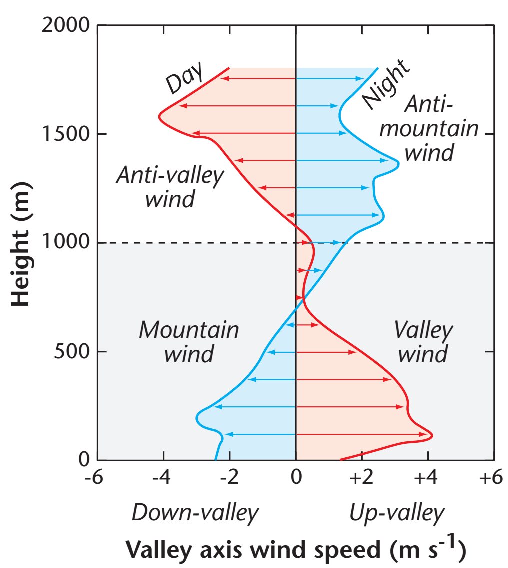

Figure 4: The vertical distribution of along- valley winds in a 1 km deep valley on Mt Rainer, Washington. Horizontal scale is graduated in units of wind speed and separated into two wind directions (up and down valley) (after Buettner, 1967).

Katabatic and mountain-wind. At night the valley surfaces cool by the emission of long-wave radiation. The lower air layers cool and slide down-slope under the influence of gravity. These *katabatic* winds usually flow gently downhill at about 2 to 3 ms-1, but greater speeds are observed where the cold layer is thicker and the slope steeper. The convergence of these slope winds at the valley centre results in a weak lifting motion (Figure 2b). All of the downslope flows combine into a down-valley flow known as the *mountain wind* which seeps out of the mountain valleys onto the adjacent lowlands. A counter flow (the *anti-mountain wind*) flows up-valley aloft (Figures 2b and 3). The drainage of cold air down-slope or down-valley often occurs as intermittent surges rather than a continuous flow. The reason for this behaviour is not certain, but it appears that the stable cold air may become retarded or blocked by obstacles in its path, until a threshold value is reached beyond which it overcomes the restraining influences and plunges forward.

Katabatic flows are commonly found over ice and snow surfaces. If the head of the valley considered above was occupied by a glacier or snowfield the additional cold air would have augmented the nocturnal down-valley wind. In fact katabatic winds are also found by day over glaciers and ice-caps. The cold ice surface gives rise to a semi-permanent temperature inversion, and cools the overlying air layer because Q_H is directed towards the surface. The cool skin of air slides down the glacier, over its snout, and cuts under the less dense valley wind which is moving up-valley. The glacier wind usually dies out within about 0.5 km because the air is slowed, by friction with the valley floor and the opposing force of the valley wind, and because it is thermally modified at its lower boundary as it advects over the warmer valley floor. Nevertheless the harshness of the microclimate associated with the wind causes vegetation to be virtually absent for about 100 metres from the snout and to be stunted or deformed for a considerably larger distance.

Similar katabatic flows exist on a larger scale on the surface and at the margins of the continental ice-caps. In this case the cold layer is up to 300 metres deep, and the wind speeds are greater than for glacier and valley winds. Winds as strong as 20 $\mathrm{m\,s^{-1}}$ are often encountered (Figure 4).

From a practical standpoint the less spectacular gentle flow ($\approx 1\,\mathrm{m\,s^{-1}}$) of cold air on good radiation nights is also important. Height differences of less than one metre may allow cold air to drain to the lowest lying portions of the landscape (e.g. hollows, basins, valleys).

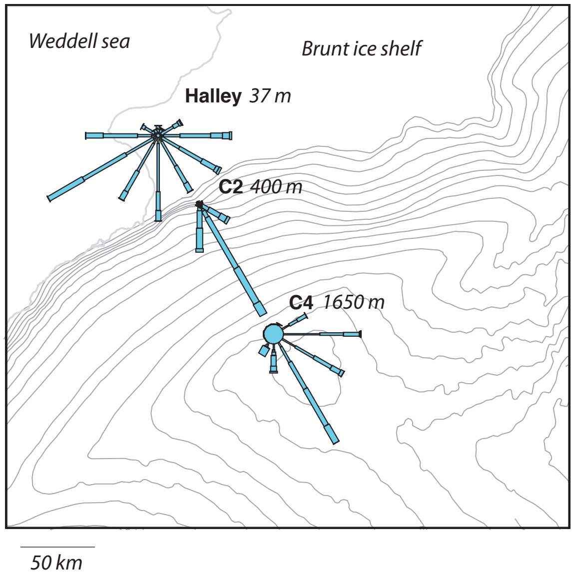

Figure 4: Katabatic flows on the continental ice slope of Antarctica. Wind roses are shown for three sites, for Halley, C2 and C4 during katabatic conditions. Plotted as a background is the 100 m topography (modified from Renfrew and Anderson, 2002)

The coldest (and densest) air settles to the lowest levels and therefore temperature increases with height above the valley floor producing a valley inversion. In this stable stratification temperature varies directly with elevation. The air draining downslope may not be colder than that already accumulated in the valley base, especially if its motion is sufficiently vigorous to produce turbulent mixing. It then over-rides the valley cold pool and may generate oscillations known as gravity waves on the top of the pool. Should cooling be sufficient to depress temperatures below the dew-point the stratification is made visible by the presence of radiation fog in the lowest-lying spots. If temperatures fall below freezing these same areas experience the greatest frost risk. Such frost 'pockets' should be avoided when planting frost-susceptible plants and trees.

Thermal belt. Under these conditions as one moves up the valley slopes from the floor the thermal conditions ameliorate, until the top of the pool of cold air is reached. Above this point the normal adiabatic decrease of temperature with height (Γ) usually prevails. Thus the most favourable location on the valley sides is just above the level to which the cold pool builds up. This is known as the *thermal belt* and its height depends upon the geometry of the valley and the area of cold air sources which feed the cold pool. The belt usually corresponds to a contour band along the valley sides which is therefore favoured for the siting of thermally sensitive crops (e.g. orchards, vineyards) and native dwellings.

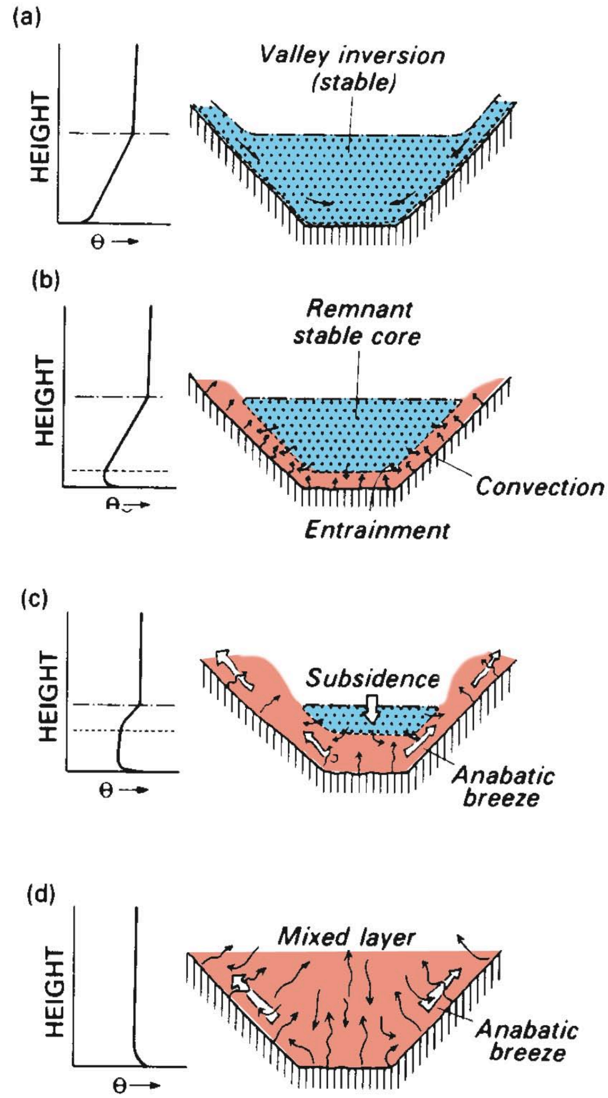

Figure 5: Time sequence of valley inversion destruction including potential temperature profile at valley centre (left) and cross-section of inversion layer and motions (right), at each time, (a) Nocturnal valley inversion, (b) start of surface warming after sunrise, (c) shrinking stable core and start of slope breezes, (d) end of inversion 35 h after sunrise. (Based on Whiteman 1982.)

The destruction of the overnight valley inversion is different to the erosion of an inversion over flat ground (see Lecture 28). When solar heating of the valley begins sensible heat flux from the surface generates a shallow mixed layer as in the flat ground case. This occurs over both the floor and sides of the valley, thereby isolating a core of stable air (Figure 5b). As these mixed layers develop thin anabatic flows up the valley sides remove mass from the core causing the elevated inversion to sink and therefore warm adiabatically (Figure 5c). Eventually the upward movement of the mixed layer and the descent of the inversion top combine to eliminate the stable core and a well-mixed atmosphere fills the valley cross-section (Figure 5d).

TOPOGRAPHICALLY MODIFIED WINDS

The airflow over non-uniform terrain is not easy to generalize. Every hill, valley, depression, tree, rock, hedge, etc. creates a perturbation in the pattern of flow, so that the detailed wind climate of every landscape is unique. It is, however, possible to isolate some typical flow patterns around specific features and we will consider some here, but if the detailed wind field is required it is probably best to model the situation by building a scale model and subjecting it to flow simulations in a wind tunnel.

Before describing wind characteristics it is helpful to discuss the concept of separation. Flow over a flat surface normally adheres to it, sometimes called the 'no slip' condition. It is possible for the flow to become separated from the surface. An adverse pressure gradient in the flow can bring it to a standstill or even cause it to reverse. This can happen as flow passes over a sudden discontinuity in the surface (e.g. over a sharp corner, or over a steep obstacle or cavity). The flow cannot fully adjust to the topography, it separates from the surface and a low pressure region is created which sucks part of the fluid towards it, often generating turbulent vortices in its lee. These lee eddies are seeded into the flow behind the discontinuity forming a highly turbulent wake. This is the way in which mechanical turbulence is generated at the surface, around each pebble, blade of grass, boulder, tree, house, etc. It also happens at the scale of topographic features if the curvature is sufficiently large. Here we classify flows around topographic features according to whether or not separation occurs.

Flow over moderate topography. The varying elevation of the surface over moderate topography (slopes up to about 0.3 (17°)) usually allows the boundary layer flow to adjust without separation. Essentially an increase in the ground elevation relative to the mean requires the flow to constrict vertically and this results in acceleration. Conversely a drop of surface elevation results in a slowing down.

Applying these basic rules to some simple topographic forms (Figure 6) it follows that in comparison with unobstructed flow ahead of the feature, and at the same height above the local surface:

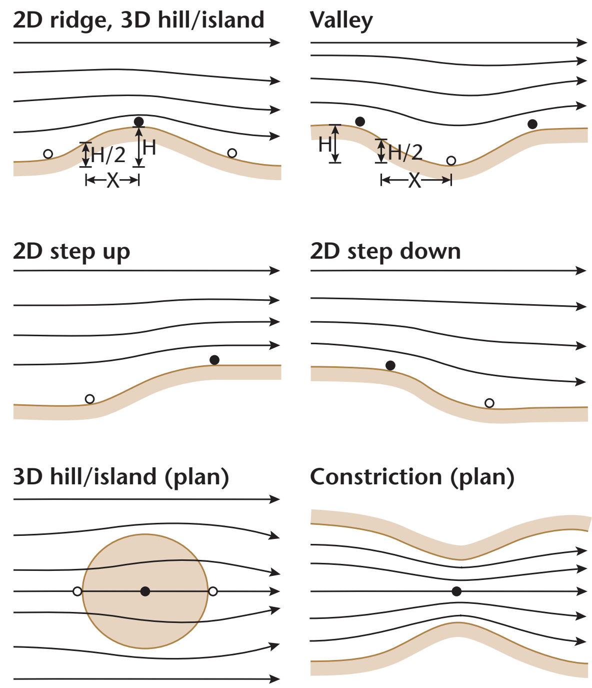

Figure 6: Typical patterns of airflow over moderate topography. The point of maximum (●) and minimum (○) is also indicated.

(a) the two-dimensional ridge will cause a speed-up in its vicinity with a maximum near the ridge crest.

(b) the valley will produce a slowing with maximum shelter near its floor.

(c, d) the step change will result in a speeding up with a maximum at its crest for flow up over it, but a slowing down with a minimum speed near the base of the slope for flow downwards.

(e) the three-dimensional hill or island will increase speeds over it, like a ridge, but also around it (see plan view) with a maximum at the summit.

(f) the valley 'neck' or mountain pass will produce a jetting through the gap with a maximum at its narrowest point (gap flow).

Taylor and Lee (1984) synthesized the results of observation and theory on these flows. They suggest that the maximum amplification factor u_{max}/u_{up} (where u_{up} is the upstream mean windspeed at the same height above its local surface as the wind, u_{max}, is above the hilltop) can be estimated using the formula:

$$u_{max}/u_{up} = 1 + b(H/X) \qquad (20.1)$$

where H is the height of the topographic feature and X is the distance from the crest of the hill or top of the step to the upstream point where the height equals H/2 (illustrated in Figure 6). The recommended value of b is 2.0 for a 2D ridge, 1.6 for a 3D hill and 0.8 for a step up. The formulae and b values also apply to the maximum diminution factor ($u_{\min}/u_{\mathrm{up}}$) for the appropriate 'inverse' form (e.g. valley instead of ridge) with the modification that H is negative. The largest amplification factors are about 2, indicating a doubling of the unobstructed wind speed. More typical values are up to 1.6 1.8. These are maximum speed-up values which would occur close to the surface above a hilltop; smaller increases are experienced in the envelope that extends from near the upstream edge of the topographic feature, up to about X above it, and a short distance downstream from the crest for a ridge or hill, but further for a step change.

It should be emphasized that these equations apply to neutral stability with flow normal to the 2D features. Amplification values will generally decrease in unstable conditions and when flow is at lesser angles to the axis of the feature. In stable conditions speed-up (slowdown) may be greater than the neutral case because the flow adjustments are restricted to a shallow layer. An elevated inversion is particularly important because the flow must squeeze through a narrow gap between the hilltop and the inversion base. It also causes more of the flow to adjust laterally, i.e. around the sides of a three-dimensional hill, because horizontal movement is less constrained.

Flow over steep topography. If the upwind or downwind slope of the ground exceeds about 17°, flow separation occurs. This is accompanied by secondary flows. Such complex systems do not lend themselves to mathematical analysis; therefore we will simply describe some common cases as shown in Figure 7.

The principles of separated flow are seen in each case:

(a) as the flow approaches the steep ridge a pressure build-up occurs. The pressure is a maximum in the middle to upper part of the face. The major portion of the flow moves upwards (to lower pressure) and 'squeezes' over the ridge with a major speed-up at the top. Some of the flow is deflected and drawn downwards (to lower pressure) and forms a bolster eddy (or roll vortex) along the base. Here flows are in the opposite direction to the general flow and winds are weak, unsteady and turbulent. As the rest of the flow streams over the ridge separation occurs at the top and a lee eddy forms. Again winds at the surface are counter to the mean flow, weak and unsteady. Hence the area is sheltered in the mean but subject to short-term gustiness. Above and slightly downstream of the ridge the conditions are conducive to convective cloud formation. For stably stratified flow, a series of recurrent, but diminishing lee waves and their associated lee wave clouds may form downstream.

(b) the general flow may skim over a narrow steep valley without much adjustment. Within the valley a single lee eddy may be created as a tangentially-driven secondary flow. Light and variable winds or semi-stagnation may be found at the bottom of the valley. a wide valley with steep walls is really a combination of a step-down and a step-up as shown in the succeeding two cases. Overall the valley produces subsidence which may be sufficient to warm the air and produce an area of preferred cloud dissipation.

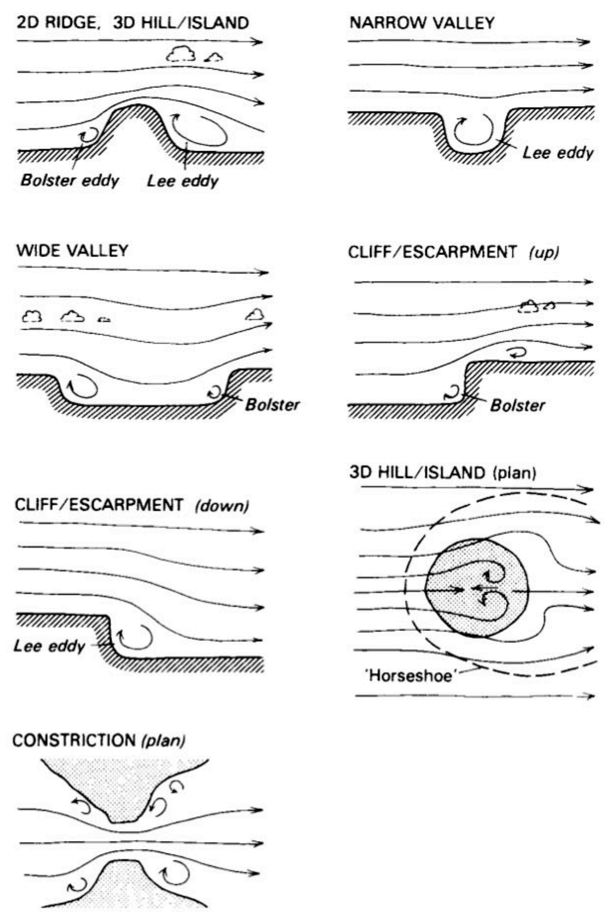

Figure 7: Typical patterns of airflow over steep topography.

(c) flow up a steep step such as a coastal cliff or an escarpment produces a bolster at the base, strong jetting over the cliff edge and often a lee eddy on the top, slightly back from the edge. It is obviously a favoured site for cloud formation.

(d) flow down over a step creates a strong lee eddy and helps to dissipate low cloud.

(e) a steep isolated hill causes the air to speed-up considerably, both on the windward face and around its sides. Separation from the top and both sides produce lee eddies which are often unsteady behind the hill, as can be seen in both the side and plan view of Figure 7. Therefore in the immediate lee of the hill flow is very complicated: the wind direction near the surface may be counter to the general flow (i.e. upslope), speeds are considerably reduced, but there is great spatial variability in turbulent activity. The turbulent wake of the hill, shaped like a horseshoe in plan, extends downstream for a considerable distance.

(f) the flow through a sharp constriction is similar to the moderate case in Figure 6 except that bolster and lee eddies are present ahead of, and after, the point of narrowing.

The examples shown in Figure 7 assume the flow to be normal to the long-axis of the two-dimensional features. For most other angles the strength and persistence of the flow features will be reduced. With parallel flow little effect is expected. The results are only for neutral stability. Separation is favoured by instability and dampened by stable conditions. Separation is very marked if the lee slope is sunlit, thereby generating anabatic breezes which augment the lee eddy.

It should be appreciated that for the most part the patterns in Figure 7 apply over a rather wide range of scales. For example, the valley results apply with little modification to ravines, road-cuts and gullies if the cross-sectional form is similar.

There are many environmental and practical implications arising from these preferred flow structures. Knowledge of localized areas of speed-up are important in the siting of: windmills for power generation, emission sources to maximize pollutant dispersal, communications towers to avoid structural failure, and forest cutting patterns to minimize windthrow due to excessive wind loading. Ability to predict areas of shelter is helpful to: minimize heat loss from houses and domestic animals, plan transportation routes to prevent buffeting of vehicles, and to avoid areas of very high snow or sand deposition. Lee eddies are semi-enclosed circulation systems and therefore poor locations for a pollutant source (Figure 8). Similarly areas of persistent downward motion are to be avoided for such uses, and are relatively dangerous places in which to land aircraft. On the other hand zones of persistent uplift are excellent areas for soaring bird flight, kiting and gliding. Knowledge of local flow patterns around islands, headlands and near cliffs is of great advantage to boaters. Indeed a grasp of the principles governing local thermal breezes and topographically-modified winds is invaluable in a wide range of outdoors activities.

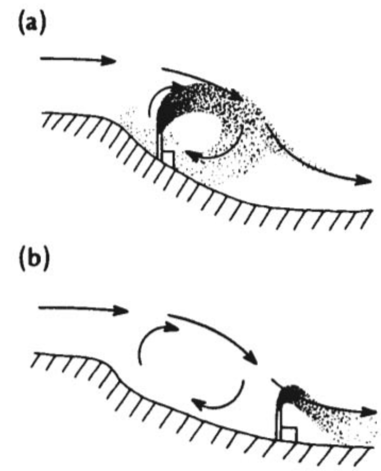

Figure 8: Problems of pollution dispersal on the windward slope of a steep-sided valley. In (a) the plume contents are trapped in the lee eddy, and in (b) are forced to ground level by 'downwash'.

Flow over roughness changes. Flows across a boundary between surfaces of different roughness induce changes in speed and direction. Figure 7 illustrates this point. The rough/smooth transitions shown could represent that between land/water, forest/grassland or urban/rural areas. The discussion assumes the location is in the Northern Hemisphere.

(a) Wind from rough to smooththe slower flow over the rough area causes the wind to be considerably backed relative to the gradient wind direction. Over the smoother area the flow accelerates and veers to the right (i.e. it becomes less backed relative to the gradient wind).

(b) Wind from smooth to roughthe transition is the reverse to (a) and as the flow decelerates it backs.

(c) Wind parallel to the boundary with the rougher area to the right of the windthe different degrees of backing relative to the gradient wind over the two areas creates a zone of air mass convergence along the discontinuity. This produces a band of stronger flow and perhaps sufficient uplift to establish a line of cloud parallel to the boundary.

(d) Wind parallel to the boundary with the smoother area to the right of windthe different degrees of backing produce air mass divergence, deceleration and cloud dissipation along the discontinuity. Hence in the belt of the mid-latitude westerly winds the most favourable weather (light winds and cloudless skies) is often experienced on southerly coasts.

(e) Airflow across an isolated area of greater roughnessthis case combines all of the foregoing effects. It can be considered to represent strong flow across an island of low relief, an isolated area of forest or a city. Applying the rules from cases (a) to (d) we see that flow. entering the rough area slows down and backs, and when exiting speeds up and veers back to its original upstream direction. Individual air trajectories would however be offset to the left compared to their original path. Convergence causes a strengthening of the flow along the left edge of the rough area, and divergence results in a slackening along the right edge. Uplift is likely to be generated especially over the left hand side of the area. Such knowledge is important in assessing trajectories of pollutants, areas of cloud modification and when seeking winds for sailing or areas of shelter.

The preceding roughness effects are expected to apply in near-neutral stability with moderate to strong winds. The direction changes would be less in unstable and greater in stable conditions. With weak or light winds thermal breezes may be present and would have to be vectorially added to determine the resultant flow.

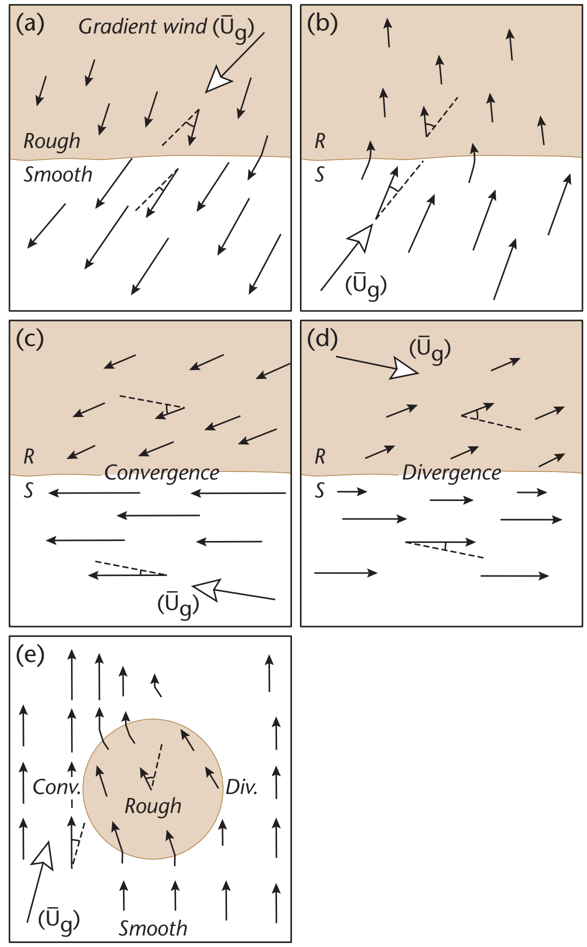

Figure 9: Effects of roughness change on wind. For details see the text. Dashed lines are parallel extensions of the gradient wind direction. Arrow lengths approximately proportional to near- surface wind speed.