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ASCOT

ATMOSPHERIC STUDIES IN COMPLEX TERRAIN

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A PLANNING GUIDE FOR FUTURE STUDIES

M. M. Orgill

Pacific Northwest Laboratory
Richland, Washington

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FOREWORD

In the Fall of 1976, a joint ERDA-EPA workshop[a] was conducted in Albuquerque, New Mexico to consider what research programs are necessary to increase understanding of transport and diffusion of pollutants in complex landforms. The workshop proposed a broad research program which would address the air quality assessment aspects of various energy conversion developments in complex terrain throughout the U.S.A. and particularly the western states. Some of the recommendations of the workshop were:

- An integrated program of mathematical model development, field experiments, physical model experiments and theoretical research involving air pollution meteorology in complex terrain should be supported for a period of 3 to 5 years.
- Several complex terrain research sites should be established as focal points of well-planned field measurements programs.

In 1978, the Assistant Secretary for the Environment, Department of Energy (DOE), developed a program aimed specifically at Atmospheric Studies in Complex Terrain (ASCOT). The ASCOT program is designed to develop the technology needed to assess atmospheric properties and the impact of new energy sources on air quality in areas of complex terrain.

The ASCOT team is composed of scientists from DOE-supported research laboratories and university programs. The initial research focus was placed on the study of nocturnal drainage winds; the initial study site was chosen in the Geysers geothermal area of Northern California. This study is presently planned to continue in 1982.

Early-on, the ASCOT participants recognized the need for program planning beyond the drainage wind study in the Geysers area. By summarizing past research, recommending future research topics and surveying potential study sites, this document provides a basis for future planning and selecting both

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(a) Barr, S., R. E. Luna, W. E. Clements and H. W. Church. 1977. Workshop on Research Needs for Atmospheric Transport and Diffusion in Complex Terrain. CONF 7609160, September 28-30, 1976, Albuquerque, New Mexico, 17 pp.

the technical aspects and location of the field experiments for relevant DOE research related to atmospheric boundary layer flow coupled with the influence and sometimes domination of underlying complex terrain.

Marvin H. Dickerson
Scientific Director
ASCOT Program

SUMMARY

This study assists the U.S. Department of Energy in conducting its atmospheric studies in complex terrain (ASCOT) by defining various complex terrain research options and relating these options to specific landforms or sites. This objective included:

- reviewing past meteorological and diffusion research on complex terrain
- relating specific terrain-induced airflow phenomena to specific landforms and time and space scales
- evaluating the technical difficulty of modeling and measuring terrain-induced airflow phenomena
- evolving several research options and proposing candidate sites for continuing and expanding field and modeling work.

To evolve research options using viable candidate sites, four areas were considered: site selection, terrain uniqueness and quantification, definition of research problems, and research plans.

APPROACH

Selecting candidate site or sites will depend on whether a particular energy development project is planned to go into operation or undergo expansion. It will also depend on whether this particular development may pose a possible environmental problem. Candidate sites will most likely be located where potential oil shale, geothermal, and coal gasification-liquefaction projects are anticipated.

A second alternative to site selection is to choose a site primarily on a generic or theoretical basis. That is, choose a location on the basis of a particular landform (such as a hill, ridge, valley, etc.) with the object of investigating in detail the dynamics, kinematics, and thermodynamics of the terrain-induced airflow and diffusion characteristics of that landform. Results from models and field measurements would then be applied to various landforms for validation.

Terrain uniqueness and quantification has not been covered extensively in the complex terrain literature. Terrain uniqueness implies that each landform has its own special terrain characteristics that in turn interacts with the lower atmosphere to produce their own peculiar topo-climate and meteorology. Terrain quantification (defining and evaluating terrain factors such as geometry and vegetation that are important to a particular site) needs more emphasis and would lead to a better understanding of how topography interacts with the atmosphere.

For several prospective sites the research problem will be defined by the particular landform or site, the proposed energy development, and the possible environmental problems caused by a certain type or combination of meteorological phenomena. In some cases the research problem may be well defined and specific and would be essentially solved once the problem is carefully investigated.

A more general approach to a complex terrain problem would be to investigate and understand the relatively less difficult phenomena (local and boundary layer phenomena) before progressing to the more difficult phenomena such as interaction of synoptic and local phenomena, eddies, turbulence and diffusion. Such a general procedure has in some respects been adopted by ASCOT.

Once a site has been selected for study and a research problem has been defined or proposed, then a specific research plan to study the problem is formulated before proceeding. The methods of study are well known: theoretical or analytical modeling, numerical modeling, field measurements, and physical modeling. A research plan could be designed for a specific site, a selection of sites, or just for model development and application. A general research plan for a large-scale study of a specific site or selection of sites should attempt to incorporate the capabilities and advantages of all of the above four research methods.

CONCLUSIONS

- A substantial data base currently exists on meteorological and diffusion phenomena in complex terrain. ASCOT should take advantage of this data base for guidance and assistance in its data collection, analysis, modeling and field program efforts.

- Data and information gaps do exist in the present complex terrain data base that provides ASCOT with an excellent opportunity to expand and update the complex terrain data base.
- A wide variety of problems in respect to topo-meteorology and diffusion in complex terrain still need to be investigated in detail for different types of landforms.
- In the next 5 to 10 years expansion of oil shale, geothermal, and coal gasification-liquefaction development projects in the western and eastern states will provide an interesting range of candidate sites for conducting topo-meteorology and complex terrain diffusion studies.

RECOMMENDATIONS

1. The airflow and diffusion characteristics of a particular landform are a result of the topographic details and terrain-induced airflow pattern. Since terrain uniqueness exists it is recommended that the ASCOT program considers studying a range of different landform types.
2. This study recommends that more emphasis should be placed on the quantification of the basic components of terrain in order to help understand the various interactions between landforms and meteorology.
3. On the basis of Recommendation (1) it is suggested that ASCOT seriously consider the option of conducting future studies at other energy development areas. Some possible current options for candidate sites are: 1) Oil Shale Resource Areas of Utah and Colorado, 2) Mono-Long Valley geothermal resources area (California), and 3) the Valles Caldera geothermal resource area (New Mexico). Other options could occur in the next five years because of new energy developments and political, economic and energy policy decisions.

4. A long-term (5 to 10 yr) but flexible research plan should be adopted by ASCOT. A principal objective of this plan is to use field programs and model development toward the construction of a "validated" operational model that can be utilized for different landforms and meteorology.

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1.0 INTRODUCTION

In 1978, Lawrence Livermore Laboratory (LLL) began direction of a multi-laboratory program to conduct atmospheric studies in complex terrain (ASCOT) for the U.S. Department of Energy (DOE). The program was to integrate field and physical modeling experiments with the development of mathematical modeling and research. The Geysers-Calistoga Geothermal Resource Area in northern California was selected as an initial site for conducting field and modeling studies. After two or three years of field experiments and model development on drainage-wind phenomena in the Geysers area, additional areas in complex terrain may be considered for further studies dealing with atmospheric transport and diffusion.

To assist in identifying areas for future studies, the Pacific Northwest Laboratory (PNL) is to help define various research options over complex terrain and to merge these options with candidate sites and landform types. Specifically, PNL is to:

- identify, review, and discuss known terrain-dominated flow and turbulence phenomena.
- discuss the importance of terrain features and resultant flow patterns on transport and dispersion processes.
- relate each terrain-induced airflow phenomena to specific landforms and time and space scales.
- rank each terrain-induced airflow phenomena according to technical difficulty regarding modeling and field measurements.
- evolve several options for extending the field experimental and modeling work of the ASCOT program based on available data.

The identification and review of terrain-dominated airflow borrows from past work relative to complex terrain, including various literature searches conducted under ASCOT (1980).[a] Specifically, airflow phenomena are discussed that result from interactions with synoptic, mesoscale and locally-derived airflows over complex terrain. Those airflow phenomena important to transport and dispersion are presented and related to specific landforms and time and space scales. However, this review is not providing definitive answers to all questions regarding transport and diffusion over complex landforms, nor is it providing proven models for assessing air quality over complex terrain. Instead, this report defines airflow phenomena over complex terrain, provides guidance for assessing and selecting tools or techniques for work in complex terrain, and presents information to help define future goals for ASCOT.

Future candidate sites are assessed in terms of the important airflow phenomena over complex terrain. Since the sites in the ASCOT program are associated with potential or expanded nonconventional energy development, geographical sites where oil shale, geothermal or coal gasification and liquefaction are planned are considered here. More conventional energy sources such as coal-, gas or oil-fired power plants are not included.

Finally, the report synthesizes the information to provide guidance for proceeding, or not proceeding, toward future defined goals. Included in the discussions are the definition and explanation of various options for continuing and expanding the efforts of the ASCOT program to other candidate field sites.

(a) This survey does not attempt to review in depth the available literature from other countries. However, an upcoming program is the GARP Mountain Subprogram (ALPEX) in the Alps of Europe (Kuettner 1980; UCAR 1980).

REFERENCES FOR SECTION 1.0

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2.0 TERRAIN-INDUCED AIRFLOW PHENOMENA

During the last several years, interest has increased in the transport, diffusion, and environmental impacts of air pollutants in regions of complex landforms such as valleys, hills and mountains. Weather modification projects in the mountainous West have shown concern over the targeting of seeding agents in orographic cloud systems for optimum augmentation of snowfall and seasonal snowcover. Similarly, industrial emissions in complex terrain and their environmental impacts on visual air quality, health, and climate have become important issues. Government and utilities are concerned about locating coal-burning generating facilities in rural areas and regions of complex landforms because of possible environmental impacts. Other types of industrial growth related to nonconventional energy development are expected to accelerate in regions of complex landforms creating additional air-quality concerns.

Among the specific air-quality concerns associated with pollutant release and transport in complex terrain are:

- Channeling of plumes
- Impact of elevated plumes on terrain surfaces
- Stagnation of air pollutants in basins and valleys
- Effects of local terrain-induced wind, such as drainage, anabatic, valley-mountain, on transport and diffusion
- Fumigation of elevated plumes or pollutant layers
- Effects of deformation, turbulence and terrain-induced eddies on plumes and dispersion.

Many or all of these problems are known to have some effect on the siting of plant locations, stack height requirements, emission controls, and on meeting the various air-quality requirements for a particular region.

Any planning study involved in selecting field sites for future work on transport and diffusion must consider landforms. In many studies the classic terminology of landforms, such as hills, valleys, cliffs, or mountains, has been used with only minor attempts to further classify or quantify the terrain. Appendix A briefly discusses some of the possible classification or quantitative systems for describing and classifying terrain. This information is useful for defining and discussing specific landforms in this study as well as for providing some initial information on terrain quantification methods in regard to airflow phenomena.

This section presents a brief survey and description of a number of airflow phenomena which are of paramount importance to the transport and diffusion problem associated with complex landforms. These airflow phenomena are: 1) airflow deformation, 2) separation and wakes, 3) gravity waves, 4) local winds, 5) stagnation and natural ventilation, 6) eddies and vorticies, 7) momentum and thermal boundary layers, 8) turbulence, and 9) forest canopy flow and exchange processes. Each of these airflow phenomena will be defined and discussed on the basis of past studies and information. Then Section 3.0 discusses the importance of these various terrain-induced airflow phenomena on the transport and diffusion processes.

2.1 AIRFLOW DEFORMATION

Airflow over flat, rather homogeneous terrain moves unimpeded except near the surface where frictional forces decrease the wind velocity. However, when the air flows over areas of more complex terrain the air deviates from this pattern or is deformed. In this section, several airflow deformations are discussed including streamline distortion, speedup and channeling.

2.1.1 Streamline Distortion

Simple and complex landforms distort the horizontal and vertical streamlines as indicated by theoretical and laboratory studies (see Figure 2.1). For example, airflow over simplified terrain-shaped obstacles such as cylinders, spheres and hemispheres, has been studied (Milne-Thompson 1960; Lamb 1932; Vallentine 1959). These studies, as well as numerous others, have provided considerable information on the gross features of streamline patterns. However, theoretical solutions in these studies are limited to nonstratified inviscid flows and only apply to studies of complex terrain flows under

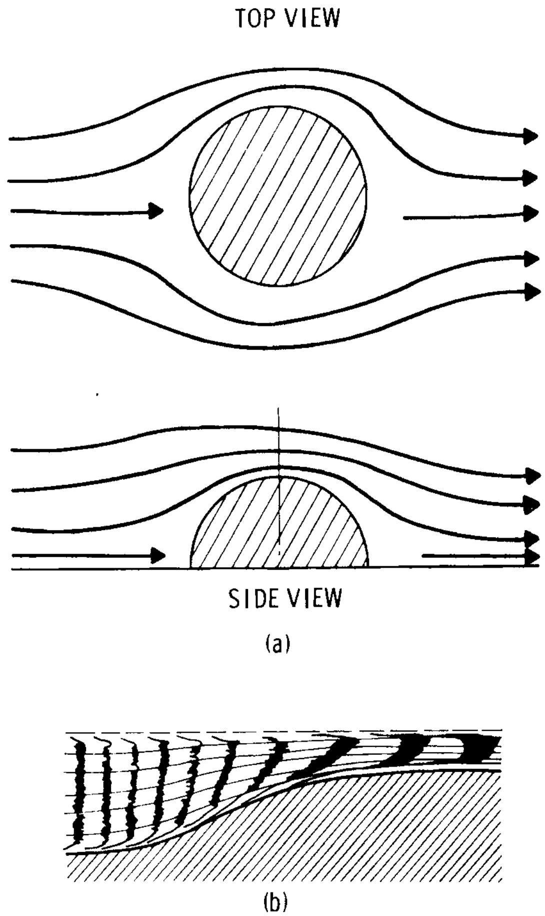


FIGURE 2.1. Airflow Over Simple Terrain-Shaped Objects: a) Top and Side Views of Potential Flow Past a Hemisphere or Idealized Hill and b) Visualization of Streamline Deformation Over a Simple Ridgelike Obstacle in a Water Channel

near-neutral or adiabatic stability conditions and where adverse pressure gradients are not likely to produce flow separation. Egan (1975) has reviewed potential flow solutions for flow over a half-cylinder (ridge) and hemisphere (hill) to examine the differences between two-dimensional and three-dimensional flow and their impact on airflow patterns and plume trajectories. His results suggest that perturbations to streamline heights above crest are, on the average, 45% smaller for the spherical shapes than for cylindrical shapes; the relative vertical separation of the streamlines over the crest is also small for the three-dimensional shapes (see Section 3 for a discussion of the implications for plume trajectories).

Mathematical characterizations of streamline distortion are generally limited to simple geometric shapes such as idealized hills, ridges, and cliffs, although landforms do have irregular geometric shapes. However, qualitative interpretation of streamlines modified by terrain as seen from field wind data is often discussed in the literature. Laboratory studies, involving wind tunnels and stratified towing tanks, have provided useful information on the gross features of streamline patterns over terrain obstacles. For example, Figure 2.1b illustrates how the streamline flow over a simple ridgelike obstacle deforms in a converging water channel. As indicated, fluid elements stretch out in the flow direction and narrow in the crossflow direction. Various wind tunnel studies have simulated streamline airflow patterns for different types of complex landforms (Halitsky 1961; Halitsky, Magoney and Halpern 1964, 1965; Garrison and Cermak 1968; Kitabayshi, Orgill and Cermak 1971; and Orgill 1971). The results of these studies are generally representative of airflow under neutral stability conditions.

2.1.2 Speed-up of Airflow

Because of their elevation, isolated hills, cliffs and ridges are generally expected to have stronger mean winds than surrounding flat or rolling terrain. However, theory and observation indicate that the airflow over a hill or ridge causes the speeding up of a sheet of intermediate fluid to velocities exceeding those of the upper layers. Thus, a lower level maximum in wind speed may develop locally. In less extreme cases, velocities may not exceed those of the upper layers but the speed may be constant, or approximately so, for some depth. This characteristic of the airflow over an obstacle is called the speedup factor (Golding 1977 and Putnam 1948) and is the result of at least three factors:

- a hill or ridge with an aerodynamic-like profile
- a barrier effect of the obstacle
- surface friction or shear stresses.

The speed-up factor [$s(z)$] is defined as the ratio of the wind speed (u) at some displacement Δz above the top of a ridge to the speed of the incident wind (u_0) at the same elevation ($h + \Delta z$) above the upwind ground level, i.e.,

$$s(z) = \frac{u(z = h + \Delta z)}{u_0(z = h + \Delta z)} \tag{2.1}$$

where h = height above ground level. Or, the fractional speed-up ratio is:

$$\Delta s(\Delta z) = \frac{\Delta u(h + \Delta z)}{u_0(h + \Delta z)} = \frac{u_1 - u_0}{u_0} \tag{2.2}$$

where u_1 is the speed at height Δz above the hill and u_0 is the upstream speed at the same height above the ground (Figure 2.2a).

If the speedup effect exists, it is normally confined to a shallow layer near a small hill or ridge summit. Davidson et al. (1964) indicated that speedup effects can occur over sand dunes and coastal areas, but are wind-direction dependent. Petterssen (1964) showed the existence of speedup factors of 1.35 for small islands off the west coast of Norway, while deBray (1973), Bowen and Lindley (1974, 1977), Freeston (1974), and Kilickaya (1980) showed that speedup effects exist for escarpments or cliffs. Others, including Putnam (1948) and Golding (1955), have summarized research on the existence of speedup.

Laboratory and theoretical models of airflow over two-dimensional ridges or escarpments generally give s(z) values of unity or greater. Frost, Maus and Fichtl (1974), Taylor and Gent (1974), and Jackson and Hunt (1975) attempted to quantify the speed-up factor as a function of height over two-dimensional isolated ridges under neutrally stable flows. Jensen and Peterson (1978), Peterson, Kristensen and Su (1976), and Bitte and Frost (1976) present over-speed estimates for small escarpments or cliffs. Mason and Sykes (1979) measured the speedup at 2 m over a 137-m hill; its value was 2.3 as compared to a theoretical estimate of 2.0. Also, Bradley (1980) measured the fractional speedup ratio on a 170-m hill in Australia and discovered an average $\Delta s = 1.07$ while the theoretical estimate was $\Delta s = 1.24$. Height of the velocity maximum was 28 m, very close to that observed. Deaves (1976) developed a mathematical method for estimating the speedup factor for arbitrarily shaped surface obstacles and compared the results with field, laboratory, and other

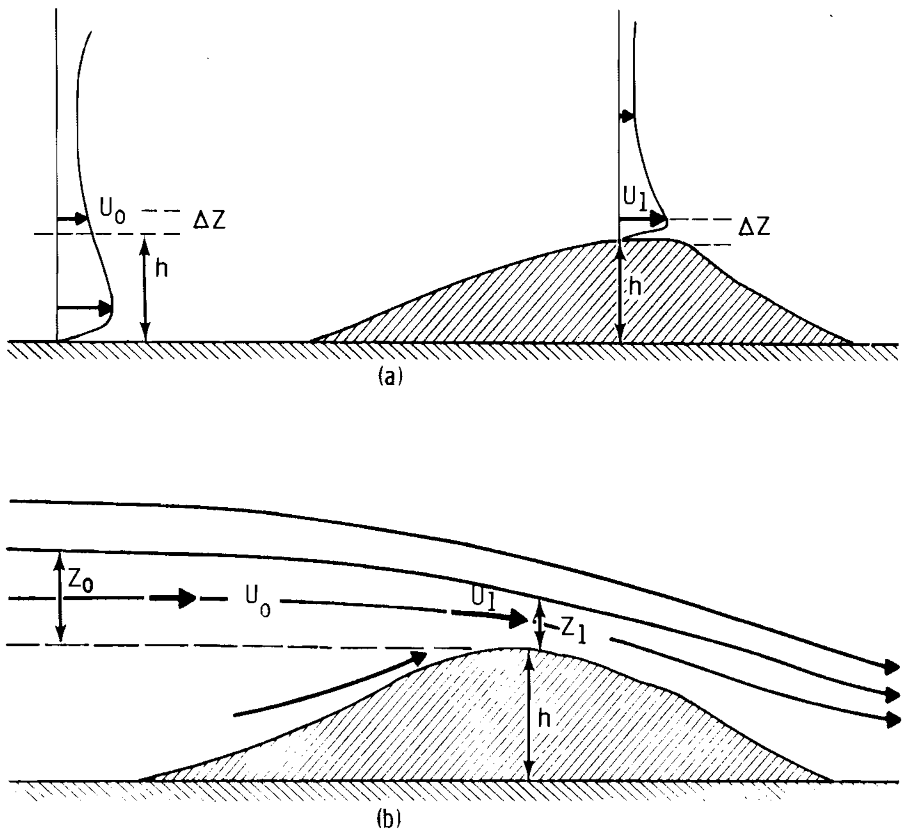


FIGURE 2.2. Pictorial Presentation of the Speedup Factor for: a) a Ridge or Hill (see Text) and b) Over a Ridge Caused by "Mounding up" of Stable Air Upstream ($U_1 > U_0$ and $Z_0 > Z_1$; U_1 Increases with Lower Z_1).

theoretical measurements. Computed velocity profiles were in reasonable agreement with experimental results over hills and embankments.

Jackson and Hunt, and Taylor and Gent found that the fractional speedup ratio reached its maximum value close to the surface at a displacement of $\Delta z \sim 4 - 8$ m. At $\Delta z \sim 2$ m, they found that $\Delta s \simeq 2h/w$, where w is the characteristic width of the ridge. For uniform potential flow over hills or ridges, $\Delta s \simeq h/w$. For $h/w > 0.05$, the speed-up factor was found to be more sensitive to the displacement above the surface of the ridge (Δz) and steepness of the hill than the fractional speedup ratio.

Current field evidence suggests that s(z) is sensitive to the geometry of the obstacle, wind direction, surface roughness and buoyancy effects, but values of unity or greater for large hills, ridges or mountains may not exist on a long-term basis. Davidson et al. (1964) indicate that speedup factors greater than unity may exist in situations where persistent elevated temperature inversions exist over ridges. In this case the mounding-up effect could result in overspeeding. Figure 2.2b shows that in the upper levels, the energy to accelerate the air over the crest is provided by a head in the form of constant pressure surfaces sloping downwind along the airstream to the crest. The slope of the constant pressure surfaces is provided by a mounding up of the air upstream of the ridge $Z_0 > Z_1$. The mounding up is the minimum height necessary for the airstream to pass the ridge. This mounding effect also results in compression of streamlines that may be accompanied by an acceleration of the air flow over the ridge. This effect would be more noticeable for ridges of considerable height, length, and width than for isolated hills or mountains. However, for large mountains, the local wind system and roughness of the mountain itself could interfere with the speed-up effect (Wahl 1966).

2.1.3 Channeling and Other Venturi Effects

Channeling of airflow refers to the orientation of the wind approximately parallel to the axis of a valley, gorge, pass or gap because of the constraining effects of bounding ridges. In many geographical settings, high wind speeds and turbulence are observed in or through the topographic constriction. Channeling or venturi effects are often responsible for local high winds such as the Tehuantepecer, Jockwinde, Dusenwind, Kossava, Wasatch, Santa Ana, Junta, Mistral and Columbia River Gorge winds (Huschke 1970).

The physical processes governing the basic channeling and venturi phenomena are generally understood but details of horizontal and vertical windfields for particular geographic settings are not well documented. The simplest concepts of mass continuity and extended Bernoulli's Theorem provide a qualitative understanding of channeling and venturi effects. Figure 2.3a represents an ideal gorge with airflow blowing towards low pressure. If Coriolis and frictional effects are neglected, the extended Bernoulli Theorem indicates a positive increase in the kinetic energy of the airstream when

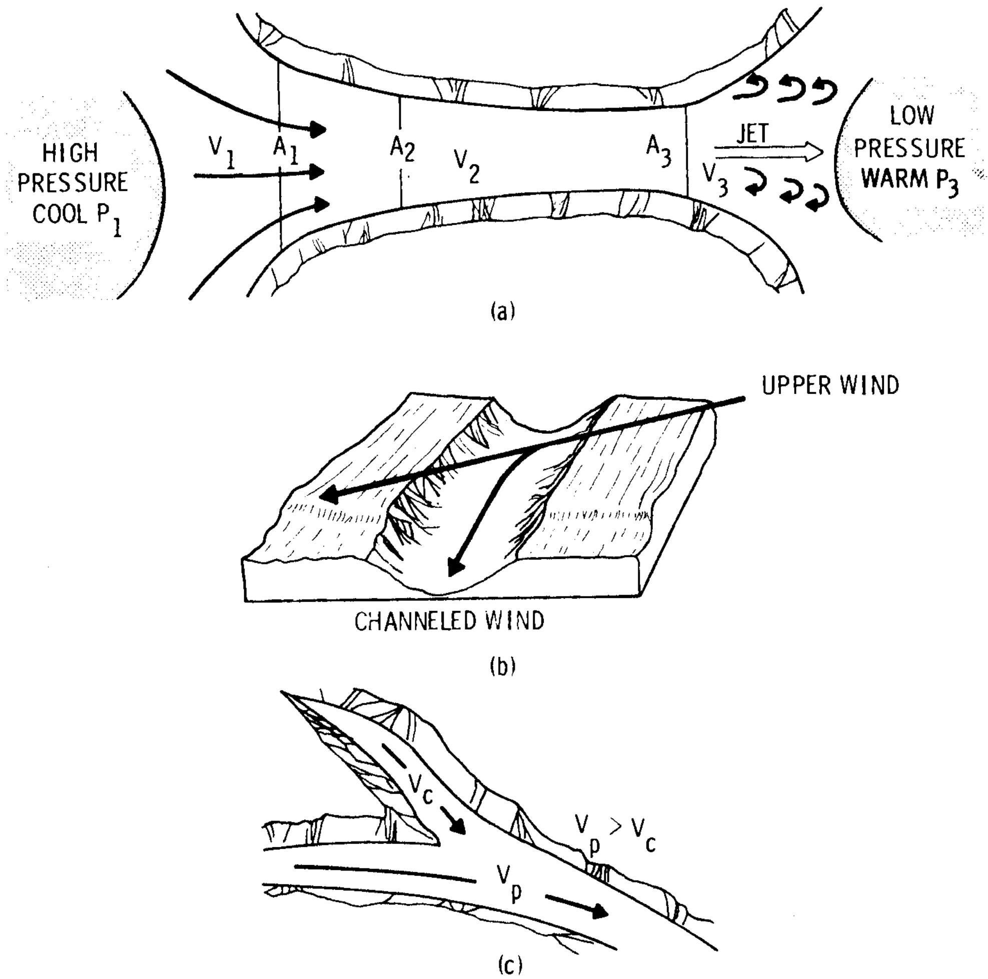


FIGURE 2.3. Possible Examples of Channeling of Airflow Because of: a) Gorge, b) Valley, and c) a Valley and Satellite Canyon ($V_p > V_c$)

$$\frac{1}{2}\ v_3^2 - v_1^2 = \frac{1}{\rho}\ P_1 - P_3 + g\ z_1 - z_3 \qquad P_1 > P_3 \quad z_1 > z_3 \qquad (2.3)$$

where V, P, g, Z, and ρ are velocity, pressure, gravity, height, and air density, respectively.

Topography can increase kinetic energy of the airstream through a gorge in at least three ways. As the airstream approaches a gorge, it converges (given the equation of continuity) resulting in

$$A_1 > A_2 > A_3 \quad \text{and} \quad V_1 < V_2 < V_3 \qquad (2.4)$$

since

$$V_2 = V_1 \frac{A_1}{A_2} \text{ and } V_3 = V_1 \frac{A_1}{A_3} \quad . \tag{2.5}$$

If a cool pool of air occupies the windward entrance to a gorge it is responsible, in the absence of strong gradients of pressure or wind above, for the excess of pressure on the windward side. The cool air "leaks" through the gorge so long as the air below mountain-top level on the lee side is warmer (Olsson, Elliott, and Hsu 1973). In addition, the height of the inversion or stable layer that caps the cool air must be lower ($z_1 > z_3$) on the lee side of the gorge (Scorer 1952). In mountain passes, the height of cool air may be increased because of the elevation difference between the pass and lower elevations.

As air converges in the narrow part of a gorge, it follows fairly smooth streamlines. When it emerges from the leeward opening, the pattern may not be the same. The air may not readily fan out to a lower velocity but emerge as a "jet" or "vena contracta," with considerable turbulence on the sides for some distance downstream (see Scorer 1952; Graham 1953; Mitchell 1956).

The preceding discussion illustrates some of the simpler aspects of channeling and venturi effects. Obviously, many variations exist because of complex topography and the effects of stability, friction, turbulence and wind direction. FIgure 2.3b illustrates qualitatively how a wind oblique to a valley or gorge axis may be channeled by the surrounding ridges. Figure 2.3c shows how air from a feeder canyon may be constricted and forced out by a faster airflow in the principal valley. Channeling and venturi effects need to be studied in detail--first through model studies (see Baines 1979) and then by field observations.

Modeling efforts indicate some of the qualitative aspects of channeled flow into valleys and gorges. For example, Anderson (1971) offers an estimation technique based on simple mass continuity. Tingle and Bjorklund (1973) calculated channeling in a broad valley using a vertically integrated two-dimensional representation of the fluid flow (the shallow fluid equations). Onishi (1969) calculated vertical wind profiles and horizontal winds over a strait using a numerical model with different roughnesses for sea and land topography. Physical models in wind tunnels have also provided qualitative and some quantitative information on channeling in valleys and gorges (Orgill 1971 and Meroney, 1979). Channeling in valleys has also been discussed in reports on field efforts (Orgill 1971, Reid 1979, Spurr 1959, and Goldman 1979).

2.1.4 Summary

Airflow deformation is one of the most obvious physical effects on airflow over or in complex terrain. It is absolutely essential that these effects are understood and modeled accurately in order to obtain reasonable wind field predictions and estimates of pollutant transport in complex landforms. A number of field and modeling efforts (such as ASCOT) will improve the necessary techniques for understanding and predicting these phenomena in the future.

2.2 SEPARATION AND WAKES

In theory, an airstream over a relatively flat uniform surface achieves a quasi-steady-state velocity profile within the friction or boundary layer. As this airstream flows over a hill or ridge the air near the surface is speeded up and the steady-state friction layer is disturbed. As a consequence, the velocity profile over the hill or ridge is modified and pressure gradients are created in the direction of the flow. These pressure gradients are also experienced by air close to the surface that is more retarded by friction than the main body of the airstream and most of the friction layer. If friction together with the adverse pressure gradient produce stagnation at any point, then reversed flow at the surface takes place on the downstream side of that point. This phenomenon is called aerodynamic separation (see Figure 2.4).

When obstacles occur in high Reynolds number flow, separation and turbulent wakes are created. The turbulent wake becomes the modified turbulence downstream of an obstacle because of random vortex shedding and subsequent cascading energy to the smaller scale components. Separation and turbulent wakes are rather common airflow phenomena in the lower atmosphere and are easily produced in laboratory airflow experiments.

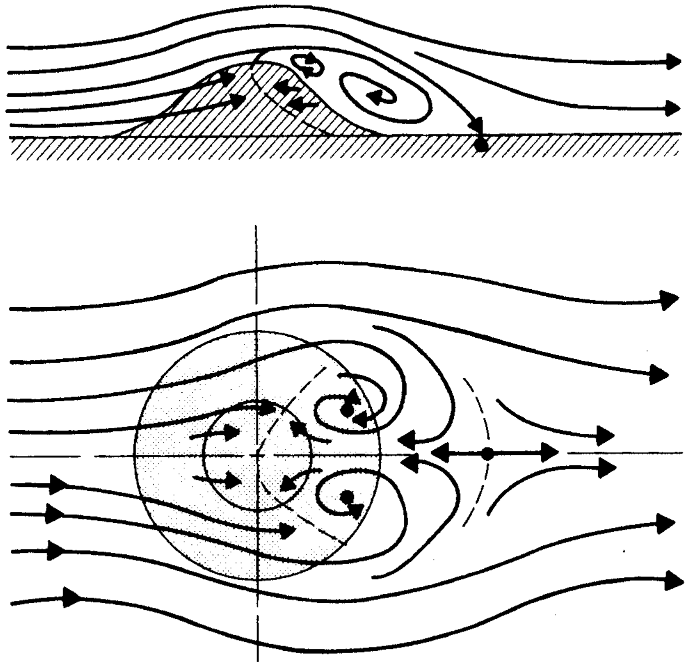


FIGURE 2.4. Aerodynamic Separation and Turbulent Wakes Downwind from an Idealized Hill

Generally, separation in an airstream over an obstacle can be classified as: two-dimensional when a wake or separation eddy is closed such as in airflow over a ridge or escarpment (cliff), or three-dimensional when the eddy is not completely closed and the air is continuously replaced, e.g., airflow around an isolated hill or mountain. These two classifications of separation will be briefly reviewed for escarpments, ridges, valleys, and isolated hills or mountains.

2.2.1 Escarpments or Cliffs

Cliffs extend vertically from a few meters to several hundred meters. In this discussion, cliffs are assumed to be of long lateral extent so the airflow is forced over the crest. The geometrical aspects of cliffs that affect airflow are slope, shape, height, and horizontal curvature. If an escarpment has a steep profile, separation may occur at the top of a windward-facing cliff, as well as at the foot of the cliff (see Figure 2.5a). These separation areas are zones of turbulent air. Observations show that the zones of turbulent eddies at the edge of the cliff may be between 2-m and 10-m deep, and extend downwind between 50-m and 200-m (see Figure 2.5b). However, the size and position of the eddy may vary as a result of wind, stability and roughness conditions (Scorer 1955).

When an airstream moves from relatively flat terrain to a cliff's edge, the region at the foot of the escarpment may be filled with an eddy, resulting from airstream separation near the top of the cliff. Under sunny conditions, thermals may originate on the lee slope and penetrate upward across the prevailing streamlines (see Figure 2.5c). A change in stability, sun insolation, or wind speed may cause the eddy to disappear and the airstream to cascade off the cliff to lower levels (Scorer 1972 and Jones 1970).

2.2.2 Ridges

Ridge geometry affecting airflow characteristics such as separation are: height, shape, slope, orientation to the wind and summit area. When the summit of a ridge is sharp, separation occurs at the salient edge (see Figure 2.6). This separation has been illustrated in laboratory tests (Plate and Lin 1965, and Arya and Shipman 1979). The separation point and the area of the turbulent wake may vary widely depending on the ridge and airstream stability. If a ridge is rounded, the point of separation may move about the ridge profile depending on the development of anabatic (slope) winds during the day (see Figure 2.7a). Separation may also occur at the summit of a flat-topped hill, especially if the windward facing slope is rather steep (see Figure 2.7b).

Separation eddies from the lee of ridges cause a quasi-periodic leeside airstream. Queney et al. (1960) and Scorer (1955 and 1978) indicate that when separation occurs just to the lee of a ridge crest, the separation eddy that is built up gradually extends the effective size of the ridge so that the position of the first lee wave is slowly shifted downwind. Eventually the eddy builds up to such a size that it is shed and a downslope flow occurs. Consequently the position of the first lee wave is suddenly shifted upwind as the eddy is carried downstream by the wind and gradually dissipated (Numbers 1 through 9 in Figure 2.8).

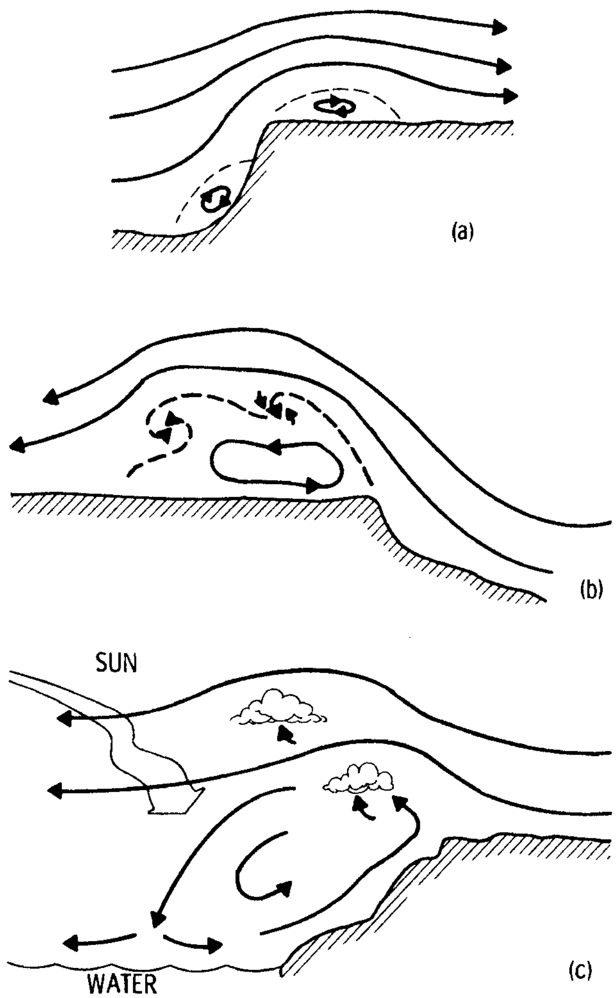


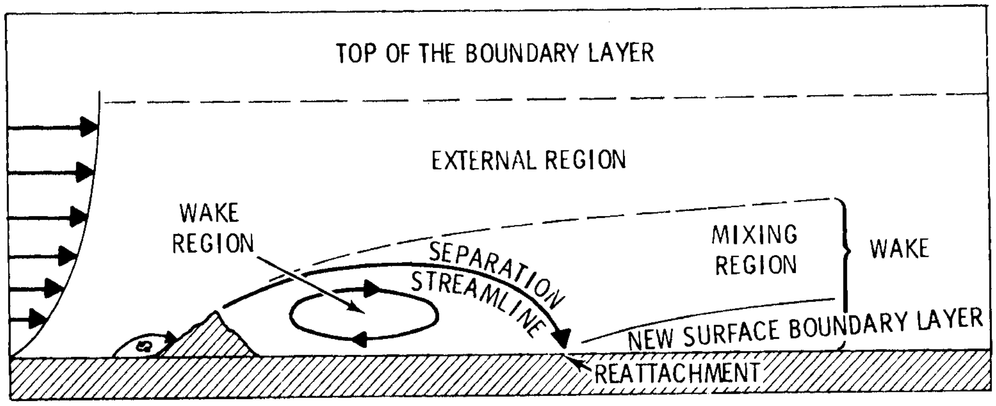
FIGURE 2.5. Examples of Flow Separation Over Cliffs (adapted from R. S. Scorer 1972)

2.2.3 Valleys and Canyons

Separation may occur in valleys and canyons, especially during periods when the prevailing wind direction is at an angle to the ridge line. In this case, the airstream separates on the lee slope and the resulting eddy may either fill the valley or a portion of it depending on airstream conditions and valley size (Figure 2.9).

Davidson (1963) obtained some qualitative observations for the distribution of vertical motion and turbulence in valleys in Vermont where the prevailing airstream was normal to the valley axis during neutral and unstable

(a)



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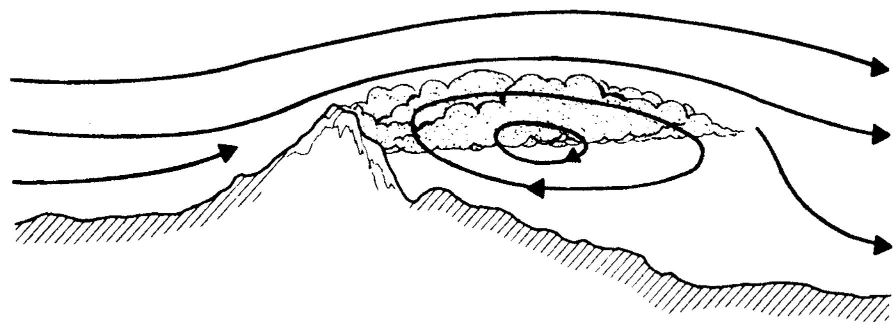


FIGURE 2.6. Schematic Representation of: a) Various Flow Zones in a Separated Boundary Layer Downwind of an Idealized Ridge in a Wind Tunnel and b) Postulated Separation for an Actual Ridge.

conditions (see Figure 2.10). According to Davidson, large leeside eddies form and dissipate in the region of maximum turbulence.

Separation occurs in the upper portion of valleys or basins when a cold airmass is trapped and becomes stagnant within valley walls (Scorer 1978). The top of the airmass is usually a cloud, the radiation from which produces a stable density continuity. Separation occurs on the leeside ridge and in combination with anabatic winds up the slope may cause an oscillation at the top of the layer. If the cold air is shallow, large variations in temperature and winds may occur on hills and ridges within the valley (Figure 2.11).

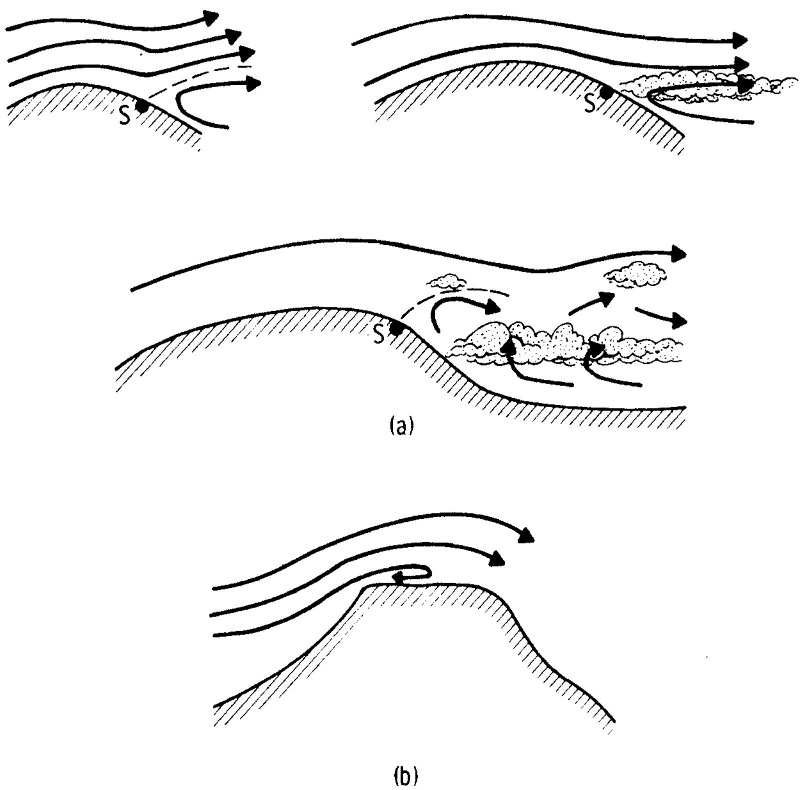


FIGURE 2.7. Movement of Point of Separation for: a) Rounded Hill in Anabatic Flow and b) a Flat-Top Hill With a Steep Windward Slope (adapted from R. S. Scorer 1972)

2.2.4. Isolated Hills and Mountains

Lee waves, katabatic winds and downdrafts from precipitating clouds can inhibit separation. Thermal convection increases separation over lee slopes but inhibits it at the top of windward slopes. Isolated hills and mountains when heated by solar isolation can become elevated heat sources, and consequently, low-lying surface air near the mountain slopes moves up the slopes to higher elevations forming thermals above the crest of the mountain. This convection favors the development of a large separation eddy to the lee of the mountain, which interacts with the wake produced by the approaching horizontal airstream (Figure 2.12). Thus, the separation phenomenon assumes a three-dimensional pattern (Scorer 1955 and 1978, and Fujita 1967).

Flow behind a three-dimensional obstacle of arbitrary shape varies when separation occurs; therefore, each occurrence has to be observed to be

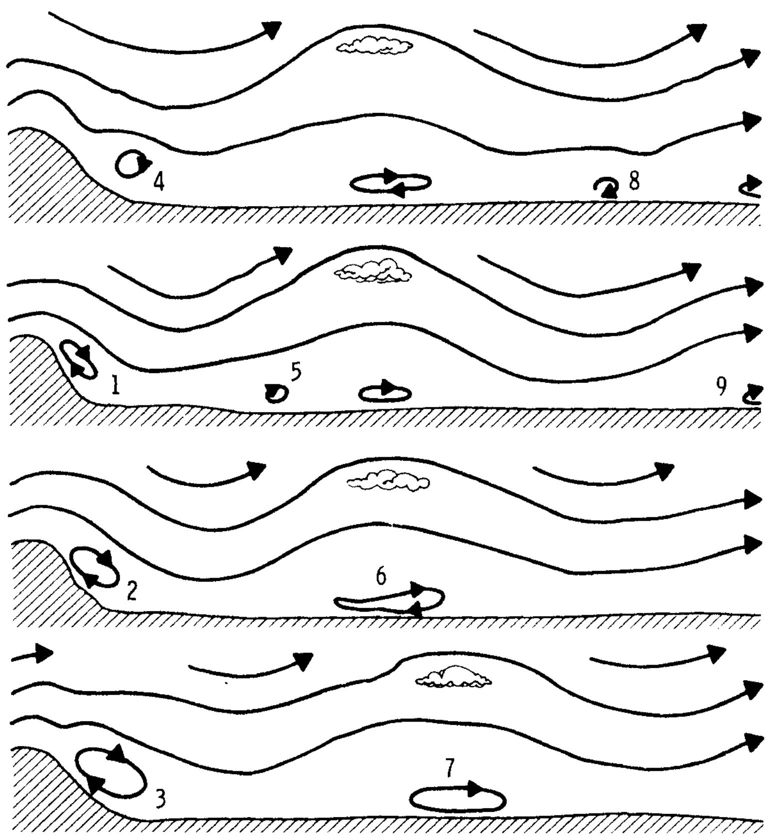


FIGURE 2.8. Periodic Release of Separation Eddies Resulting in Position Changes of Lee Waves (adapted from R. S. Scorer 1955)

discussed. For instance, Figure 2.13 shows an example of separation on a steep mountain peak as deduced from photographic observations. In this case, the prevailing winds are relatively strong and separation is occurring behind the windward wall of the cirque (Thomann 1975).

2.2.5 Model Studies of Separation and Wakes

Separated airflows have been modeled and observed extensively under laboratory conditions. Airflow, separation, and turbulent wakes have been simulated over model hills consisting of hemispheres, triangular and sinusoidal ridges and other simulated terrain. (See Hansen and Cermak 1975; Hawthorne and Martin 1955; Cermak, Lin and Chiu 1970; Plate and Lin 1965; Halitsky 1961; Halitsky, Magony and Halpen 1964 and 1965; Orgill, Cermak and Grant 1971; Kitabayshi, Orgill and Cermak 1971; Garrison and Cermak 1968; and Hunt and Synder 1980.)

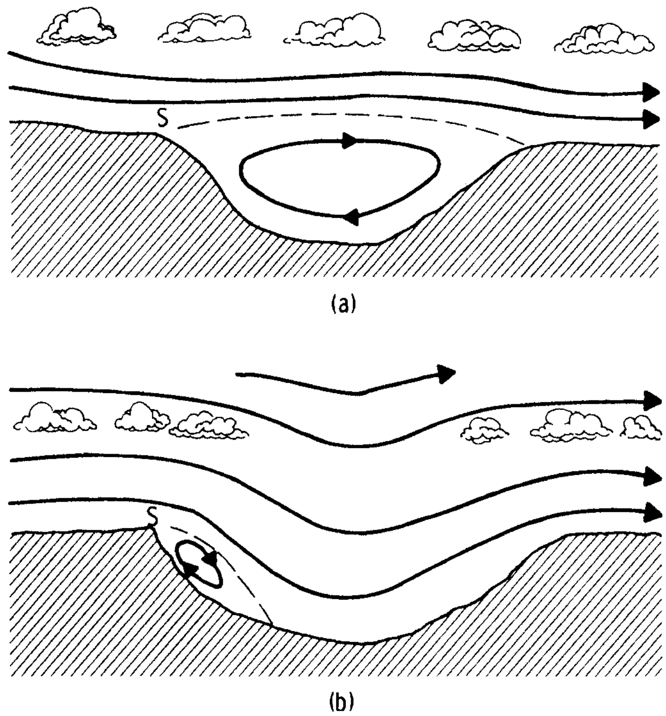


FIGURE 2.9. Examples of Valley Eddies: a) Eddy Filling a Valley Caused by Lee-Slope Separation and b) a Small Eddy Produced by a Salient Edge of a Valley (adapted from R. S. Scorer 1972)

Recently, Stewartson (1974) developed the triple deck theory that incorporates an interaction between the boundary layer and the free stream permitting the description of strong perturbations. Smith (1973), Hunt (1971), and Sykes (1978) are among the first to apply the theory to flow over obstacles. Mason and Sykes (1979) have compared the triple deck theory to solutions using two-dimensional numerical integrations of the Naviers-Stokes equations for flow of an Ekman layer over a ridge of different height scales. In these studies, upstream and downstream separations were observed to depend on the height of ridge, boundary layer depth, and stability. Hirt, Nichols and Romero (1975) have numerically simulated separated flow behind a hogback ridge using a computer model.

2.2.6 Summary

The phenomena of separation is difficult to study and quantify because of its sporadic occurrence and behavior. Separation is not easily measured in

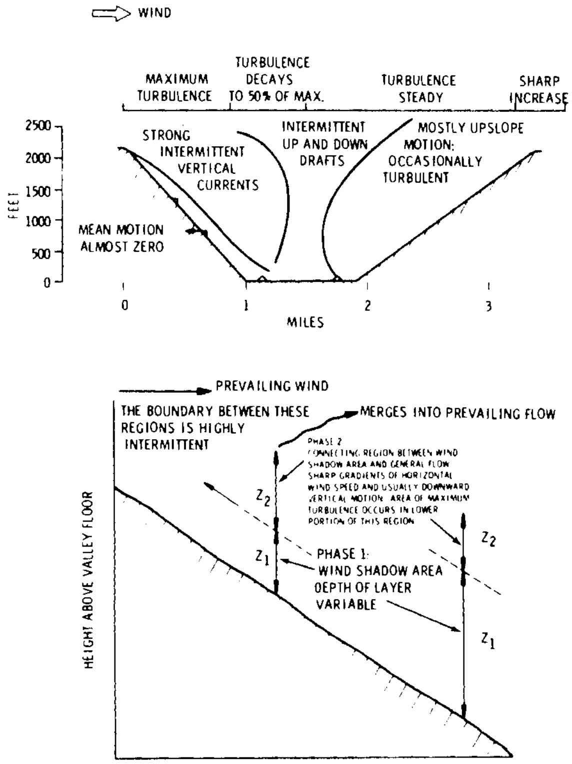


FIGURE 2.10. Distribution of Vertical Currents and Turbulence Within an Idealized Valley During Neutral and Unstable Conditions with a Prevailing Wind Normal to the Valley (adapted from B. Davidson 1963)

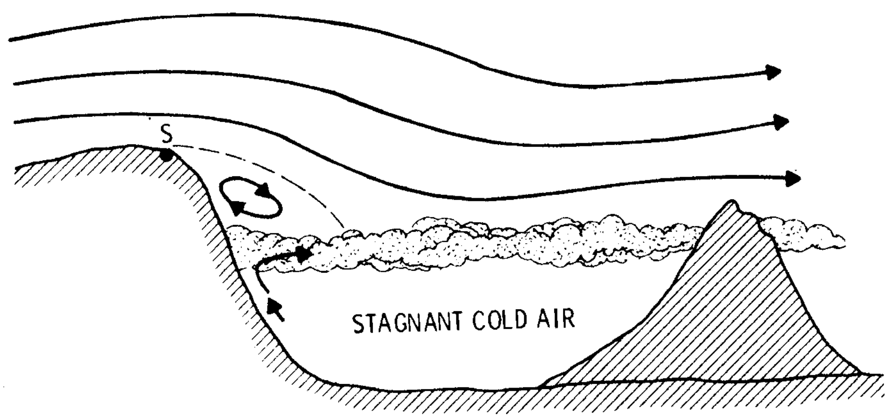


FIGURE 2.11. Illustration on how the Behavior of Separation Eddies on the Lee Side of a Valley Ridge May Affect the Position of Billows or Other Disturbances on an Elevated Cloud Layer

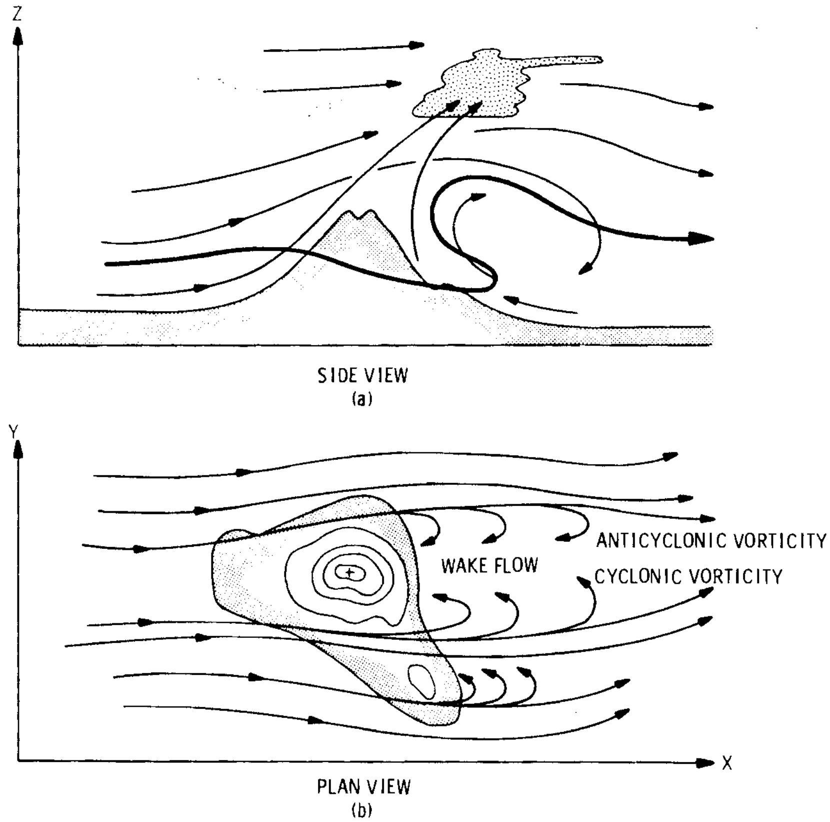


FIGURE 2.12. Effects of a Large Isolated Mountain on Airflow when Heated by Solar Insolation (adapted from R. S. Scorer 1955 and T. Fujita 1967)

the field and modeling techniques have just started simulating some of its general characteristics. Physical modeling has simulated separation for many years but there has been little effort in making a systematic study of the charcteristics of separation for different landforms. Although the general aspects of the generation and dissipation of separation are well known many detailed aspects of the phenomena such as persistence, size of eddies, interactions with ambient winds, local weather effects, need to be quantified for several different landforms.

FIGURE 2.13. Example of Separation on Eiger Nordwand (Alps) Made Visible by a Cloud

2.3 GRAVITY WAVES

Waves in a stratified flow are a promiment and dramatic feature over and to the lee of a mountain range that stands perpendicular to the prevailing winds, such as the Sierra Nevada in California, the Southern Alps in New Zealand, or the Andes in South America. These waves are basically gravity waves, since they result from the buoyant force of gravity in a stably stratified fluid and are excited by flow over an obstacle in such a fluid.

Waves generated by a fixed obstacle in a stratified flow or by a moving body in a stratified fluid at rest can roughly be classified into five flow regimes depending on the values for the following parameters (Miles 1968):

$$\varepsilon = h/\ell \tag{2.7}$$

where h is the height of the obstacle and 1 a characteristic lateral scale (ℓ = W/2 where W is width of the obstacle) and

$$Nh/U^{(a)}$$

where U is a characteristic fluid or airflow velocity and N is the intrinsic (Brunt-Vaisala) frequency

$$N = \frac{-g}{\rho_0}\frac{d\rho}{dz}^{-1/2} \tag{2.8}$$

The five flow regimes are illustrated in Figure 2.14 and are defined as follows:

- Laminar or potential flow represents the lower limit of the regime where Nh/U $\rightarrow$ 0.
- Lee waves represent the range 0.30 < Nh/U < 1.00.

(a) This parameter is the reciprocal of the Russell number (Ru = U/Nh) or a densimetric Froude number.

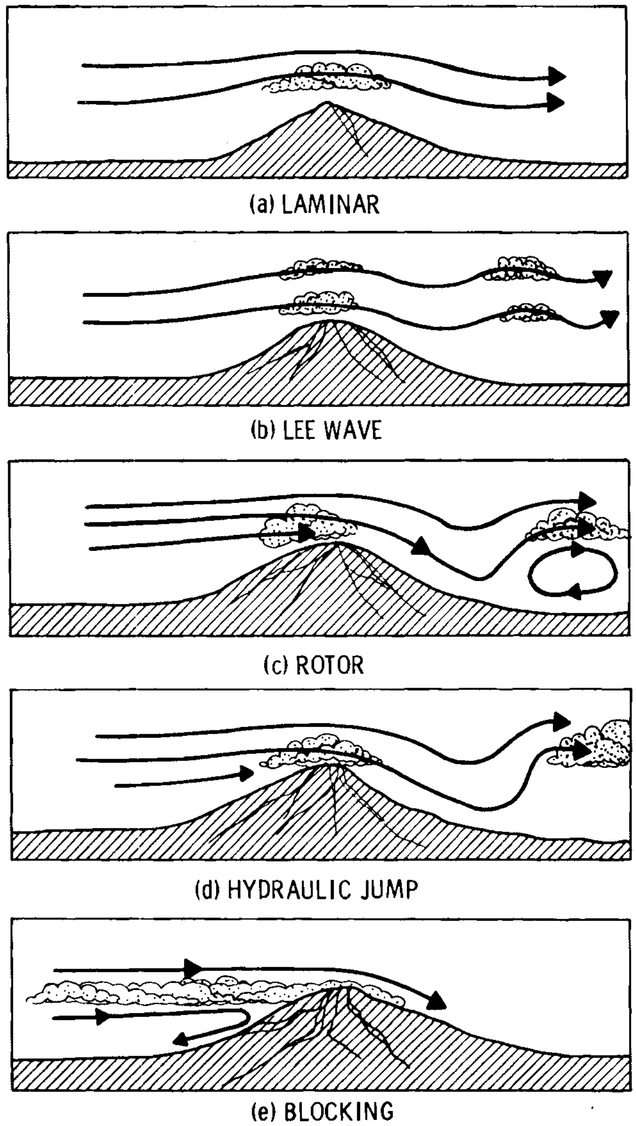


FIGURE 2.14. Five Flow Regimes That Can Occur with Stratified Flow Moving Over a Ridge or Isolated Hill or Mountain

- Rotor regime represents the upper limit of the lee-wave regime marked by flow reversals and closed streamlines.
- Hydraulic jump occurs as Nh/U increases above unity subsequent to changes in the flow leading to formation of horizontal jets.
- The topographic blocking regime occurs as Nh/U becomes > >1 or $\rightarrow \infty$ and the flow upstream becomes completely or partially blocked.

Since the above classification is primarily based on fluid mechanical and experimental data, its usefulness in the real world of gravity waves is limited. Nevertheless, the classes serve as a means for briefly discussing some of the principal aspects of stratified flow over complex terrain.

2.3.1 Laminar Airflow

In the laminar regime, the airflow comprises a smooth shallow wave over the obstacle or ridge, providing only small vertical currents and no downstream phenomena. Waves in the immediate neighborhood of mountain ridges are called mountain waves to distinguish them from lee waves that are found 20 km to 300 km downwind of a ridge. Streamlines over the ridge (or ridges) may not have quite the same shape as the ground profile but will resemble it. Depending upon the moisture concentration, a wave cloud may form over the ridge. The cloud will form a flat base and an arched top above the condensation level (Scorer 1972).

2.3.2 Lee Waves

A complex assortment of lee wave characteristics is possible depending on atmospheric stability, wind conditions, and landform shape. The commonest type of lee wave has an increase in amplitude upward from the ground to a maximum somewhere in the lower troposphere, but then it decreases at higher levels. The general requirements for lee waves are (Queney et al. 1960):

- Long mountain ranges produce lee waves better than isolated hills or mountains.
- Steep, pronounced lee slopes are more favorable for lee waves than gentle slopes.
- A concave curvature of a ridge crest line toward the oncoming airstream is more favorable for waves than a convex shape.
- The interference between the flow above particular mountains and lee waves caused by other landforms upstream produces damping and amplification of waves.

- To produce waves, marked stability of the airmass must occur where the airstream is disturbed by the ridge, i.e., within several hundred meters of their crest level, although the stability should be less at higher levels.
- Onset of nocturnal cooling often favors the development of waves while insolational heating during the day inhibits wave activity.
- Usually, the wind must be within 30 degrees of the perpendicular to the ridge to produce lee waves.
- Wind speeds at the crest level of 7 ms^{-1} are favorable for lee waves over small mountains, rising to about 15 ms^{-1} for large mountain ranges (i.e., Rockies, Sierra Nevada).
- The airstream should comprise a deep current of air in which wind direction changes little with height and the speed increases upwards through the troposphere to produce waves.

One of the basic theoretical requirements for lee waves to occur is that the following equation have a maximum in the lower or middle troposphere,

$$\ell^2 \equiv F(z) = \frac{g\beta}{\bar{u}_o^{\,2}} - \frac{1}{\bar{u}}\frac{d^2\bar{u}}{d\,z^2} \tag{2.9}$$

where β equals $\theta^{-1}\ \delta\theta/\delta z$, and θ is potential temperature (a measure of the static stability) u_0 is the wind component perpendicular to the mountain ridge at height z in the undisturbed airstream. Generally, the second term of ℓ^2 is neglected, although it is important in any marked shearing layer. The vertical variations of F(z) in Equation 2.9 is important in determining or forecasting the existence of lee waves (Scorer 1949 and Queney et al. 1960).

Two important characteristics of lee waves are wavelength and amplitude. In theory, wavelength is much easier to calculate than the amplitude because the wavelength of any lee wave downstream of a ridge directly depends on the characteristics of the airstream. The wavelength lies somewhere between the maximum and minimum values of $2\pi/\sqrt{F(z)}$ or $2\pi\bar{u}_0/(g\beta)^{1/2}$. Since the mean stability through the whole troposphere does not vary widely, the lee wavelength is approximately proportional to the mean tropospheric wind speed or

$$\lambda = 1/2\ \bar{u}_T \tag{2.10}$$

where λ is in km and $\bar{u}_T$ in ms^{-1}. Empirically, Corby (1957) derived the following regression relation:

$$\lambda(km) = 0.585\bar{u}(ms^{-1}) - 2.8 \tag{2.11}$$

Symmetrical ridge theory indicates that the crest of the first lee wave downstream of a mountain ridge is often less than one wavelength from the mountain crest and the spacing is $3/4\lambda$.

The amplitude of lee waves depends on both the topography and airstream properties. For symmetrical mountain ridges, the amplitude depends on the height of the ridge above the surrounding countryside and also on the horizontal scale of the ridge. The dependence on height is expected, while the dependence on horizontal scale is a resonance effect. If the horizontal scale of the ridge roughly coincides with the lee wavelength, the wave amplitude will be much larger than for both broader and narrower mountains. Another topographic factor favoring large amplitude waves is the occurrence of steep lee slopes.

Lee wave amplitudes depend on profiles of wind and temperature. The largest amplitude waves occur when the airstream satisfies the condition for waves by only a small margin. In this region, large changes in amplitude may result from small changes in the airstream. Apart from this sensitive region, larger amplitude waves are more likely in airstreams containing a shallow layer of greater stability than in conditions of lesser stability through a deep layer.

Lee waves in terms of amplitude have been studied as: small amplitude waves, large amplitude waves, and three-dimensional waves. A vast amount of literature on small amplitude lee waves is available (see Corby 1954, Queney et al. 1960, Nicholls 1973, and Scorer 1978). Large amplitude lee waves in relation to strong downslope winds, hydraulic jumps, rotors and topographic amplification have been studied, but three-dimensional lee waves that occur when waves are generated by isolated mountains have not been studied as extensively as the other lee waves.

A general mathematical treatment of three-dimensional lee waves is involved (see Kochin 1938, Scorer and Wilkinson 1956, Wurtele 1957, Palm 1958, Crapper 1962, Blumen and McGregor 1976, and Gjevik and Marthinsen 1978). Brighton (1978) and Hunt and Snyder (1980) reported on an experimental study with simple geometric forms. Kochin (1938), Scorer and Wilkinson (1956), and Crapper (1962) show that the lee-wave pattern is similar to that produced by a ship moving across the surface of deep water. The waves are confined to a wedge-shaped region behind the wave source. The half-angle of the wedge depends upon the airstream characteristics and is approximately 15 degrees to 20 degrees. These waves are of two types: transverse waves with the wave crest nearly perpendicular to the wind direction and diverging waves that radiate out from the wake. The diverging wave type may be rather common for isolated islands (Gjevik and Marthinsen 1978).

2.3.3 Rotors

Theory and observations show that wave amplitude can be as large as, or even larger than, the ridge height. The largest lee waves occur when the airstream flows down a high mountain plateau without separation from the surface or where several parallel ridges are spaced at a whole number of lee wavelengths apart (Figure 2.15). Stationary lee waves produced by a large, high mountain ridge such as the Sierra Nevada and Rockies, often break up into turbulent longitudinal whirls or "rotors" in the lower layers of the airflow. These rotors are the regions containing flow in the opposite direction to the main stream with large amplitude waves.

Mountain lee waves associated with rotors are normally associated with the lower turbulent zone (Lester and Fingerhut 1974). The lower turbulent zone is a highly turbulent region of nearly neutral stability found immediately to the

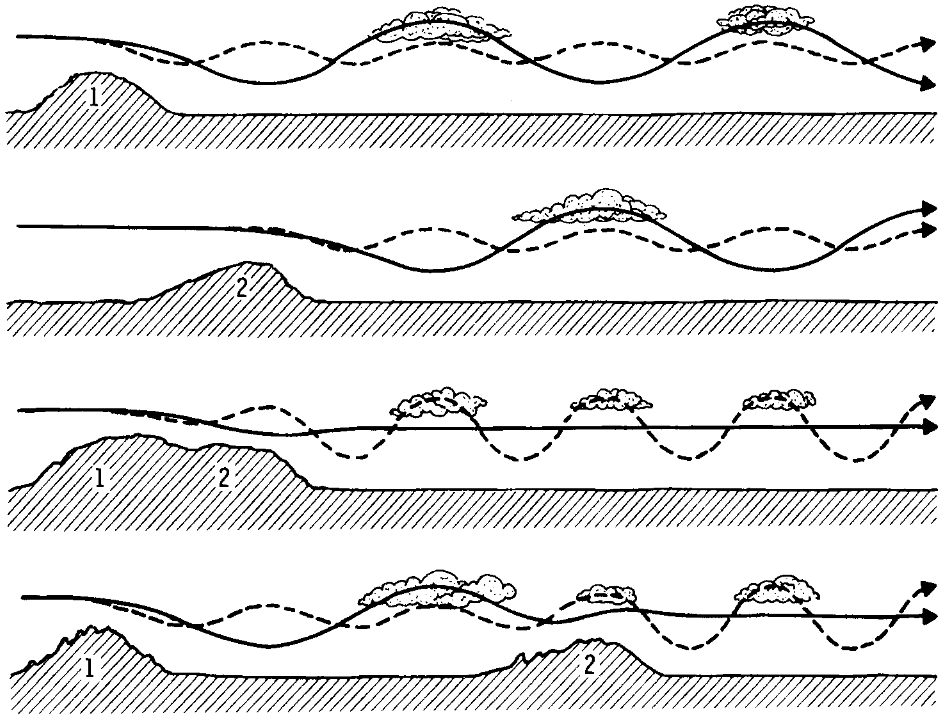


FIGURE 2.15. Effect of the Topographic Obstacle Position on the Wave Length of Lee Waves: Mountain 1 Produces Lee Waves Shown for Two Air Streams One Having Twice the Wavelength of the Other. When Mountain 2 is added the lee waves may be increased or decreased in wavelength (adapted from Scorer 1972)

lee of the mountain between the ground and an elevated stable layer in which the wave motion is occurring (Figure 2.16). Lester and Fingerhut (1974) indicate that the horizontal dimensions of the lower turbulent zone vary from 25 km to more than 65 km downstream of the first lee wave trough. Vertical dimensions range from a few hundred meters above ground level (AGL) at the lee wave troughs to 3 km at the wave crests. Turbulence levels vary from light to severe over more than 90 of the zone. Severe turbulence is commonly encountered near the upstream side of the rotor, where the largest horizontal and vertical wind and temperature gradients are found.

Kuettner (1939 and 1959), Förchtgott (1949), and Gerbier and Berenger (1957) have studied many rotor features in the lower turbulence zone. A classic field study on lee waves and rotors associated with the Sierra Nevadas was documented by Holmboe and Klieforth (1954 and 1957). Other studies have been

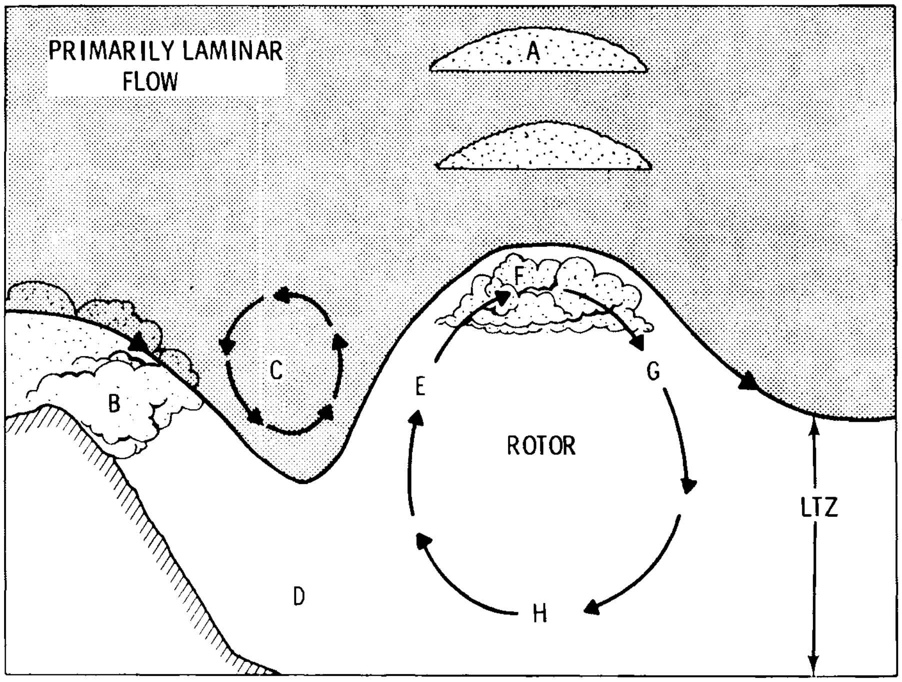


FIGURE 2.16. Idealized Cross-Section of the LTZ: a) Lenticular Clouds, b) Cap Cloud, c) Reversed Rotor, d) Region of Gusty surface Winds, e) Region of Strong Updraft and Extreme Turbulence, f) Roll cloud, g) Region of Strong Downdraft and Severe Turbulence, h) Lower Position of Rotor Circulation and Occasionally Reversed Surface Winds (adapted from Lester and Fingerhut 1974)

done in Colorado (Kuettner and Lilly 1968, Lilly et al. 1971, and Lilly and Zipser 1972). Long (1953 and 1959) and others have reproduced rotor phenomena under laboratory conditions. Scorer and Klieforth (1959) and Queney (1955) have examined the rotor problem theoretically. Their results indicate that rotors are caused in part by two factors: 1) waves near the surface reaching a critical amplitude and transforming to vortices and 2) situations where the ambient wind is very small at ground level then rapidly increasing with altitude.

2.3.4 Hydraulic Jumps

Experiments (Long 1953, 1954, 1955, 1959) with stratified flows and simulated landforms show that flow behavior depends on the shape and height of obstacles and the Froude number:

$$Fr = \frac{U}{Nh} \tag{2.12}$$

where U is the fluid velocity and Nh the phase velocity of gravity waves. Long and others have shown that for a certain range of Froude numbers a flow phenomenon resembling a hydraulic jump on a running stream may be observed in an experimental channel flow. This hydraulic jump is a steady-state, finite amplitude disturbance in which a fluid passes turbulently from a region of (uniform) low depth and high velocity to a region of (uniform) high depth and low velocity.

Houghton and Kasahara (1968) used the one-dimensional time-dependent shallow water equations to define the approximate range of conditions for hydraulic jumps. The criteria for hydraulic jumps, as presented in Figure 2.17, use two parameters:

$$Fr \equiv \frac{U_o}{\sqrt{gh_o}} \text{ and } M_c = \frac{H_c}{h_o} \tag{2.13}$$

where U_o and h_o are the velocity and height, respectively, of the approaching flow far from the obstacle (ridge), and H_c is the height of the obstacle crest. The flows characterized by the parameters F_o and M_c in Domain I in Figure 2.17 are subcritical with the free surface (interface height) dipping symmetrically over the obstacle. On the other hand, the flows in Domain III are supercritical, with the free surface rising symmetrically over the obstacle. In Domain II, jumps occur on both sides of the ridge. The jump on the windward side of the ridge runs upstream. The jump on the lee side moves downstream in Domain IIb and remains stationary over the lee slope of the ridge in Domain IIa. Domain IV corresponds to the total blocked-flow region.

A leeside jump can occur for a relatively low Froude number of the flow upstream of a ridge. This event may partially explain the not infrequent occurrence of strong winds (e.g., chinooks, foehn) along the leeside of several mountain ridges and singular mountains (Beran 1967, Brinkmann 1974, and Klemp

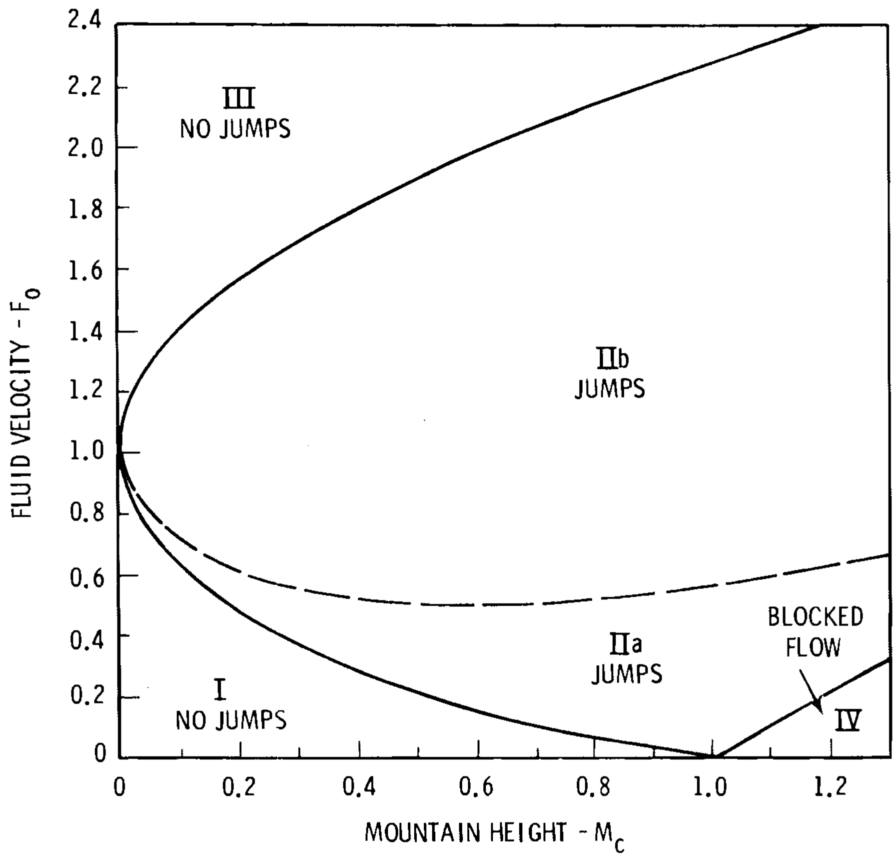


FIGURE 2.17. Classification of Asymptotic Flow Conditions (hydraulic jumps) as a Function of the Initial Flow Speed or Froude Number (F_0) and Maximum Height of a Ridge (M_c) (adapted from Houghton and Kasahara 1968)

and Lilly 1975). In the atmosphere, favorable conditions for jumps and strong downslope winds occur when a shallow upstream layer bounded by a temperature inversion near the ridge crest flows across the crest and becomes significantly depressed as the flow accelerates down the lee slope. As the flow accelerates the Froude number becomes supercritical. In some cases the downstream conditions require subcritical flow but the resulting lee waves cannot disperse the energy fast enough for a smooth transition from supercritical to subcritical. As the amplitudes grow and wavelengths become shorter, the waves break and create an abrupt transition back to a subcritical Froude number. The resulting jump has local pressure changes, turbulence and in some cases, convective cloudiness. A jump can occur on a slope when a supercritical flow moves onto a horizontal plane and must decelerate to subcritical (Turner 1973).

Theoretical and numerical models characterizing some aspects of atmospheric hydraulic jumps have been derived by Houghton and Kasahara (1968); Tingle and Bjorklund (1973); Oobayashi (1970); and Chao, Chuang, and Yan (1966). However, the nature and frequency of occurrence of jumps on smaller terrain features such as valley or hilly slopes remain to be documented.

2.3.5 Topographic Blocking

An obvious difference between airflow over flat or gently rolling terrain and airflow approaching large topographic barriers is the natural blocking or dam effect of the terrain. This blocking effect refers to the total upstream phenomenon that is the direct or indirect result of the physical obstacles presence. First, the type of landform and geometric characteristics (height, length, width, shape, horizontal curvature) determine the extent of the phenomenon. High mountain ridges several kilometers in length are more favorable for blocking phenomenon than isolated hills or mountains. Second, the static stability of the atmosphere, height of the boundary layer and the vertical profile of the wind (wind direction and speed) can reinforce or weaken the characteristics of blocking. The terrain, its local environment and the blocked airflow can also interact to produce feedback effects.

Blocking is most apparent when stably stratified air flows over high two-dimensional mountain ridges. However, issues concerning the nature of stratified flows over obstacles and the influence of an obstacle on upstream airflow are controversial. Long's experimental work (1955, 1959, and 1970) showed that for large obstacles and low Froude numbers ($Fr < 1/\pi$) blocking effects propogate upstream causing alternate maximum and minimum of horizontal velocity in the vertical. The theoretical work of Wong and Kao (1970) showed that upstream disturbances are present whenever lee waves are present, and they alter the velocity and density vertical profiles upstream. Benjamin (1970) showed that upstream blocking must be present whenever lee waves are present. Baines' theoretical and experimental work (1977) showed that upstream motion has a complex structure; however, the most important feature is: as the velocity decreases, the fluid below the top of the obstacle becomes partially blocked. This partial blocking results in rising motion and the elevated fluid forms a horizontal jet that flows down the leeside of the obstacle. Kao's (1965) theoretical studies showed that at low Froude numbers the height of the upstream stagnant zone is the same as the height of the obstacle, but as the Froude number increases, the upstream stagnant zones have been found to decrease in height for wedge-shaped obstacles.

Theoretical and model studies have indicated some of the possible upstream effects from obstacles, although questions exist regarding the representativeness of these studies. Nevertheless, these studies and observations show at least four upstream effects (see Figure 2.18).

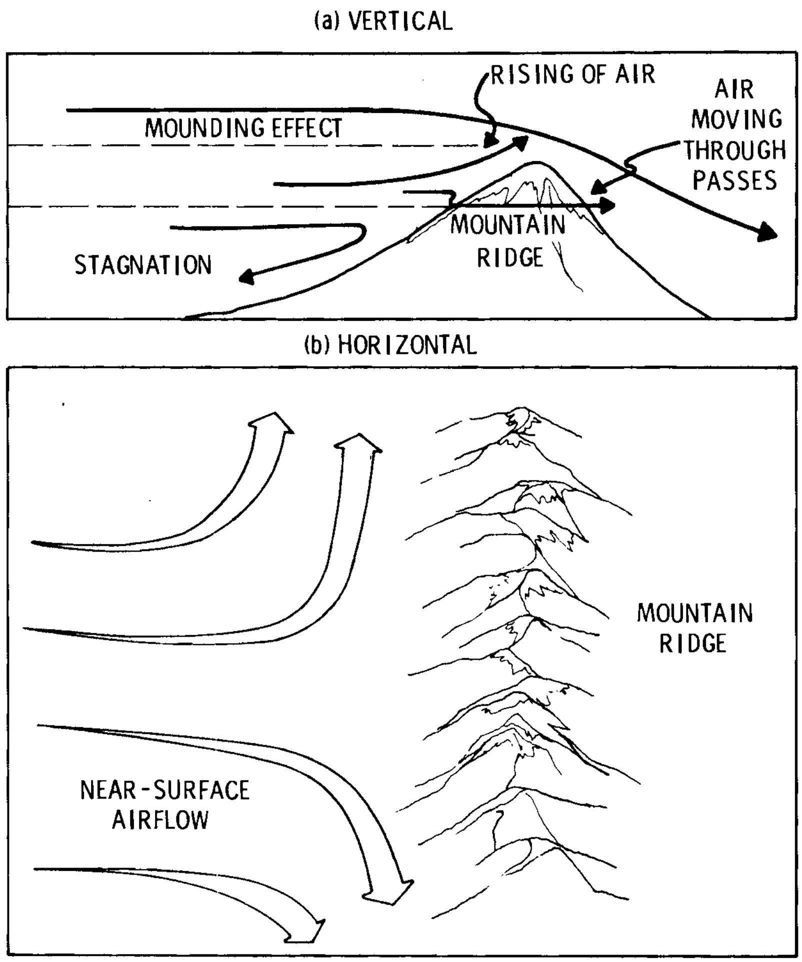


FIGURE 2.18. Vertical and Horizontal Illustrations of Various Upstream Effects Caused by Blocking topography

1. The barrier effect of a ridge may result in the mounding up of the air upstream of the ridge (Myers 1962 and Long 1970).

2. A portion of the stable air is lifted over the obstacle (i.e., the well-known topographic or orographic forced ascent of an airflow).

3. The portion of airflow lower than the height of the ridge may stagnate upstream or separate near the foothills. Ridges with gaps, valleys, and passes may allow airflow to escape to the leeside, which weakens upstream stagnation. Conditions favorable for upstream stagnation are:

   - Low-speed prevailing winds
   - Low height of the boundary layer compared to that of the ridge
   - Stable atmospheric static stability.

4. Horizontal deformation occurs upstream from a mountain ridge as horizontal airflow approaches it. Several synoptic studies in relation to the Sierra Nevadas, Cascades, and other mountain ranges have revealed this common upstream feature (Frenzel 1962).

2.3.6 Summary

Gravity waves, particularly lee waves, have been studied extensively during the past 30 years by means of theoretical, numerical, field, and laboratory techniques. As a result, several good surveys of this literature are available such as, Nicholls (1973), Scorer (1978) , and Smith (1979). The reader should consult these references for obtaining more detailed information on gravity waves and associated phenomena.

2.4 LOCAL WINDS

Local winds are defined as diurnal wind systems, restricted to relatively small areas in the range of 100 km or less. These wind systems include land and sea (lake) breezes, thermal-slope winds, and valley-mountain winds.

- Land-Sea (Lake) Breezes (LSLB) represent the complete cycle of daily local winds occurring on sea or lake coasts because of differences in the surface temperature of land and sea. The land breeze system blows from land to sea when the sea surface is warmer than the adjacent land and the sea breeze blows in the opposite direction when the sea surface is colder than the adjacent land.

- Thermal-slope winds are local daily winds that flow up a slope during the day (anabatic winds) and down a slope at night (katabatic winds). They arise, in part, from the temperature difference of the air near a slope and that of the free air at the same level in a valley or over an adjacent plain.
- Valley-mountain winds blow along the main axis of a valley or canyon, up the valley during the day and down the valley at night.[(a)] These winds result, in part, from temperature differences of the air in a valley and the much larger body of air over the adjacent plain into which the valley opens or in a wider portion of the valley downstream.

These local wind systems are better developed during the summer months than the winter months and are typically a fair weather phenomenon. They approximate the ideal diurnal pattern more closely on sunny days (clear nights) when the synoptic pressure gradient is weak.

2.4.1 Land-Sea (Lake) Breezes

The land-sea (lake) breezes are a mesoscale phenomena observed at coastal or lake shorelines and islands during periods of fair weather. This circulation system may differ considerably in character from one region to another, varying in direction, intensity, and time according to local, seasonal, and synoptic conditions. However, the prevailing large-scale synoptic pressure pattern has a direct effect on the strength, persistence, and penetration of the land-sea breezes. Various field studie- have been conducted at coastal, lake and island locations on the land-sea breezes in the United States (Table 2.1). [See Defant (1951) for the basic theory of land-sea breezes and information from other countries and earlier studies.]

(a) Davidson and Rao (1963) called this the valley-plain wind.

TABLE 2.1. Some Field Studies on the Land-Sea (lake) Breeze Wind System

State	Location	Reference
California	Point Conception to Morro Bay	Roberts (1970)
	San Francisco Bay, Lake Berryessa Area	Fosberg and Schroeder (1966)
	Monterey Bay, Salinas Valley	Read (1971a,b)
	Santa Clara River Valley	Edinger (1963)
	Los Angeles Basin	Kauper (1960)
	San Fernando Valley	Edinger and Helvey (1961)
	San Francisco Bay Area (Patterson Pass)	Porch et al. (1979)
Florida	Santa Rosa Island	Sonu et al. (1973)
Georgia	Coastal Southeastern States	Williams (1974)
Hawaii	Hawaiian Islands	Leopold (1949)
Illinois	Chicago to Milwaukee (Lake Michigan)	Lyons (1975)
Michigan	Muskegon-Holland-Grand Rapids (Lake Michigan)	Moroz (1967)
New York	New York City	Frizzola and Fisher (1963)
	Rochester-Niagra Falls Area (Lake Ontario)	Estoque et al. (1976)
Oregon	North-Central Oregon Coast South Washington Coast	Halpern (1974)
	Central Oregon Coast	Johnson and O'Brien (1973)
	Central Oregon Coast	O'Brien and Pillsbury (1974)
	Central Oregon Coast	Cramer and Lynott (1961)
	Central Oregon Coast	Burt et al. (1974)
Rhode Island	South Rhode Island Coast and Block Island	Fisher (1960)
Texas	Sabine Lake-Galveston Bay Area	Yu and Wagner (1970) Eddy (1966) Feit (1969)
Washington	Puget Sound Area	Staley (1957)

On warm, clear summer days, the horizontal temperature difference between land and water provides the energy needed to develop the vertical and horizontal air currents called sea breezes. The intensity of the daytime sea breeze normally surpasses that of the nocturnal land breeze because of the greater daytime arc of the sun during summer and the increased instability and vertical turbulence in daytime. The direction of the sea breeze is not constant during the course of the day, and this breeze often sets in with considerable gustiness sometimes in the form of a micro-cold front.

In general, sea breezes near a flat coastline blow onshore at right angles to the isotherms and isobars with speeds of 5 ms^{-1} and are limited in the vertical to about 1 km. Under the influence of the Coriolis, frictional, inertial, and pressure gradient forces, sea breezes rotate clockwise (sometimes counterclockwise) in the northern hemisphere, each rotation being completed in 24 hours in accordance with the diurnal temperature cycle. Usually, the sea breeze starts between 1000 and 1100 hours, reaches its maximum velocity between 1300 and 1400 hours, and subsides between 1400 and 2000 hours. After 1400 and 2000 hours, the sea breeze is replaced by the nocturnal land breeze. The sea-breeze fetch over water ranges up to 100 km in tropical regions but remains about 50 km in most subtropical regions. The landward penetration of the sea breeze may be around 60 km. The seaward range of the land breeze seldom exceeds 50 km as it is much weaker than the sea breeze because of terrain roughness and stabilization.

According to Schroeder et al. (1967), three types of sea breeze are reported in the literature. The first is the classical air mass front in which the sea-breeze front is characterized by a sharp fall in temperature, a rise in humidity, and a sharp change in wind velocity. The second type occurs at the wind shift line where the sea-breeze front is an airmass that is thermally modified but not kinematically altered. The third type is characterized by a substantial cooling and rise in humidity but with no wind-shear line. The inland movement of the sea-breeze front is strongly controlled by terrain features, as a sea breeze is channeled through natural openings and effectively blocked by low mountains.

A complete theory of land-sea breezes must consider: the gradient force caused by land-water temperature difference, the influence of vertical heat exchange, the turbulent friction of the air motion, the influence of the earth's rotation, the interaction of the prevailing general airflow and the topography. Most theories and models such as those of Estoque (1961 and 1962), Fisher (1961), Pearce (1955), Neumann and Mahrer (1971, 1973, 1974, 1975), Walsh (1974), Pielke (1974), Sheih and Moroz (1975), and Mak and Walsh (1976), are based on uncomplicated terrain. Fosberg and Schroeder (1966), Schroeder et al. (1967), and Mahrer and Pielke (1977) indicated that temperature, humidity, and wind patterns of the sea breeze are modified extensively as it penetrates in and over complex landforms.

McPherson (1970) studied the modification of the sea breeze circulation using a three-dimensional nonlinear numerical model. The model used a simple large bay incorporated into the surface boundary conditions. The presence of a bay produced landward distortion of the sea breeze convergence zone and within the zone definite extremes of vertical motion occurred, the positions of which were closely related to the bay configuration. The intensity of convergence and upward motion within the zone are distributed asymmetrically with respect to the bay. This distribution resulted from the Coriolis acceleration force acting in concert on one side of the bay and in opposition on the other side.

Read (1971a,b) conducted field studies at the land-sea-air interface of Monterey Bay, California, and showed that marine air penetrates at least 64 km into the Salinas Valley in west-central California. Similarly Lyons (1975) made several studies of the land-lake breezes near Lake Michigan. In general, lake breezes occurred if the following conditions were satisfied: very light gradient winds; strong insolation, i.e., usually less than 60% middle and high cloud cover during the day; and daytime air temperatures rising above lake-surface temperatures. From this, Lyons established criteria for a lake breeze (see Figure 2.19):

- The winds shift from offshore to onshore during the day.

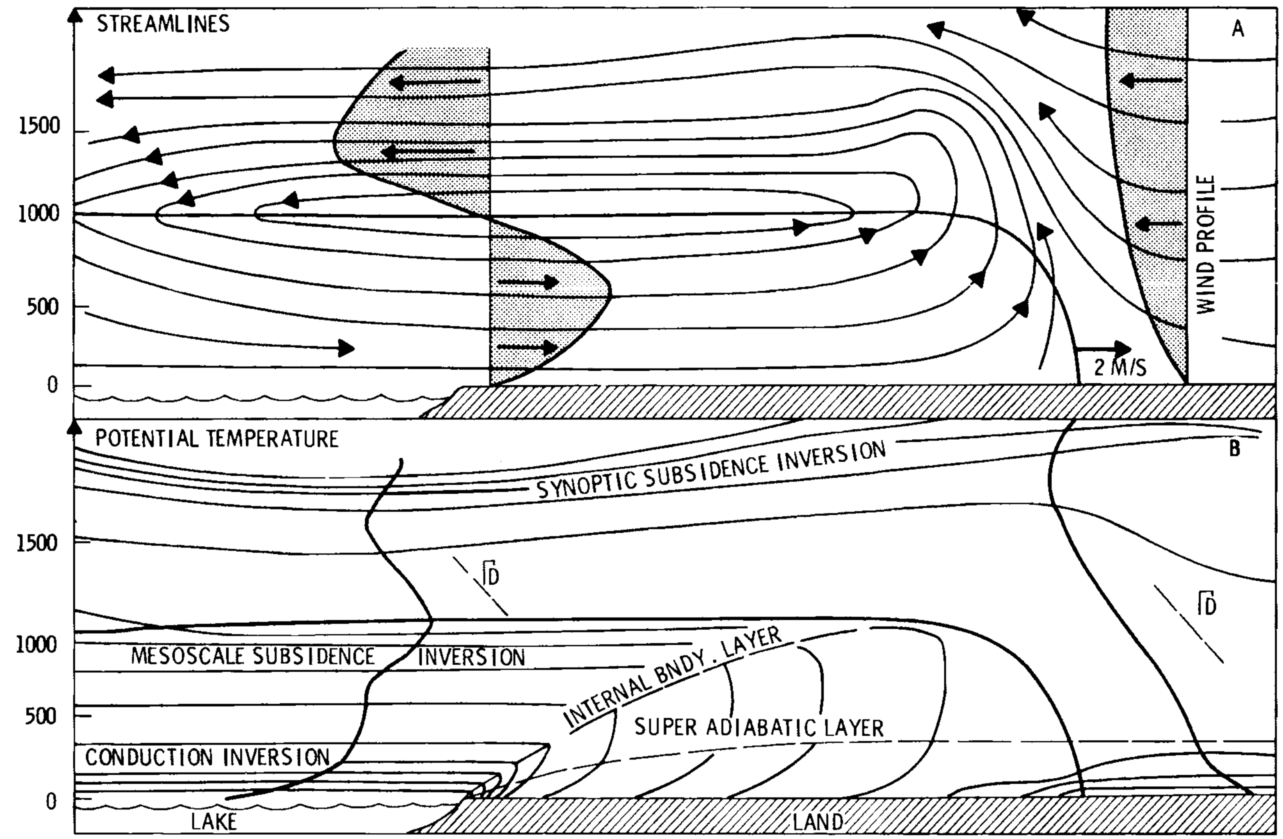


FIGURE 2.19. Typical Streamline and Potential Temperature Patterns in a Well-Developed Lake-Breeze Cell. Heavy line represents boundary of inflow (adapted from Lyons 1975)

- A definite "front" or convergence zone separates surface air streams with overlake and overland trajectories.
- A "return flow" layer aloft is discernible.

A large lake like one of the Great Lakes can have lake-breezes that have a profound influence on mixing depths, low-level winds and atmospheric stability. A large lake in complex terrain would undoubtedly exert a similar influence although no detailed documentation appears on lake effects and wind and temperature patterns in complex terrain (see Section 3.2.2).

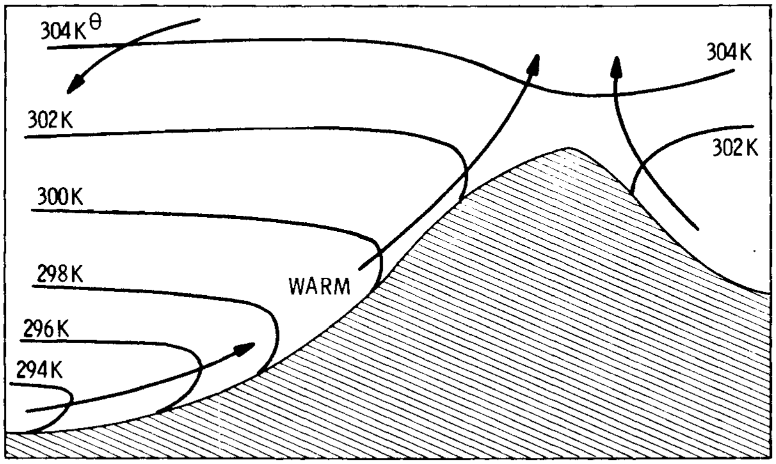
2.4.2 Thermal-slope Winds

Usually, thermally produced slope winds have an up-slope (anabatic) wind during the day and a down-slope (katabatic) wind at night (see Figure 2.20).

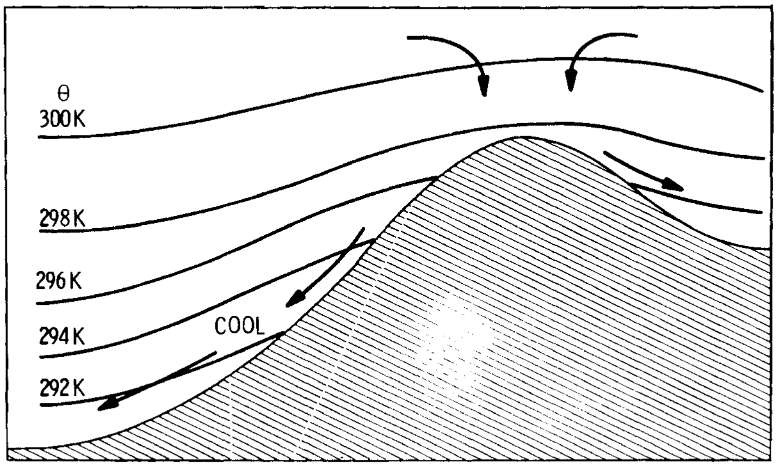
2.4.2.1 Upslope (Anabatic) Winds

During the day, elevated terrain such as hills, ridges, and mountains may act as high-level heat and moisture sources. When a mountain slope is heated by solar radiation, the air along the slope becomes warmer than the air over a valley or plain at the same elevation. A typical observed difference in mountain-valley temperature is 3°C or more (MacCready 1955). Such temperature difference creates a buoyancy force upward along the slope resulting in an upslope wind. Buoyancy effects may also result from water vapor in the atmosphere from evaporation at the slope surface or the advection of more humid air along the slope. This difference results in a more moist and lighter upslope wind than the valley or plain air at the same level. Projecting slopes or ridge summits cause the current to detach from the slope, thus increasing vertical movement or thermals (Braham and Draginis 1960).

Upslope winds start about an hour after sunrise and reach their greatest intensity at the time of maximum insolation. Because of the stronger insolation, upslope winds are well developed on the southern slope but are weaker or almost nonexistent on the northern slope. These winds prefer ravines usually of eroded slopes and are hardly noticeable on projecting ridges. The thickness of the slope wind layer varies between 100 m and 300 m, depending on wind intensity, but increases in the direction of uphill flow.



SURFACE HEATING - UPSLOPE FLOW (ANABATIC)



RADIATION COOLING - DOWNSLOPE (KATABATIC)

FIGURE 2.20. Classical Pattern of Thermally Produced Slope Winds: Upslope (anabatic) Wind and Downslope (katabatic) Wind

On the average, the slope wind speeds in the direction of the slope are about 2 to 4 ms^{-1}, their highest velocity occurring between 10 to 50 m. As the wind moves up the slope it rapidly decreases in velocity or is supplanted by other wind conditions. Since the wind current is extremely sensitive to changes in the insolation, a temporary shading of the slopes will cause an immediate response by the wind. Often the wind is weakened at the time of maximum temperature because of cloud shadows on the slopes.

Observational and theoretical studies of the upslope wind have been conducted in conjunction with the mountain-valley wind system (see Buettner and Thyer 1966 for a summary of early research). Observational studies have been conducted by Braham and Draginis (1960), Silverman (1960), MacCready (1955), Mendonca (1969), Buettner and Thyer (1966), and Fosberg (1969). Davidson and Rao (1963) found that daytime upslope winds were not observed in several valleys in southwestern Vermont, possibly because of vegetation, exposure of slopes, excessive moisture in the ground and too much wind at ridge level. Numerical or theoretical models of upslope winds have been attempted by Prandtl (1942), Orville (1964), Thyer (1966), Fosberg (1967), Gutman and Koronatova (1957) and Konyakhina, Shaposhnikora, and Gutman (1965). Generally, these studies have been successful in reproducing some aspects of upslope winds.

Closely allied to the daytime upslope wind is the cross-valley wind produced by the temperature differences in the air on opposite slopes of a valley. Hewson and Gill (1944), and Borisova and Burman (1960) observed this phenomenon; and Ekhart (1934) and Hewson and Longley (1944) discussed it. Gleeson (1951) and Tang (1976) developed theoretical solutions to cross-valley winds and illustrated how variations of friction, inertia, latitude, season, inclination of valley slopes, valley orientation and prevailing airflow affect the cross-valley wind. Their theories were compared with observations taken in the Columbia River Valley near the United States-Canadian border and in valleys of Vermont. Results were in approximate agreement with observations.

2.4.2.2 Drainage (katabatic) Wind

For downslope drainage winds to develop, two basic conditions are needed: an inclined surface, such as hill or mountain slope, and frequent clear, dry nights with radiational cooling. Drainage winds are gravitationally induced flows on a sloping surface caused by surface cooling. They generally occur at night; however, they can occur in the day over iceor snow-covered surfaces.

Buettner and Thyer (1966) surveyed the records of the early observations of drainage wind systems. As with the upslope winds, drainage winds have been studied in conjunction with the mountain-valley wind system. A few observational studies of drainage wind systems were made by Cross (1950), Snyder

(1951), Berg (1952), Wilkins (1956), Brown (1961), Bergen (1969), McNider (1977), and Manins and Sawford (1979). Brown observed the occurrence of downslope winds in the area of Fort Huachuca, Arizona. His results showed that the horizontal structure of the late afternoon winter downslope wind had characteristically abrupt windshifts, increased speed, and a maximum $\Delta T/\Delta t$ following the leading edge of the downslope wind. However, no distinct pressure change accompanied the system. Bergen observed the nocturnal (downslope) cold airflow on a small forested slope in an attempt to relate the volume and velocity of flow to the net radiation balance of the slope. In his study, the slope had an approximate 30% grade and was about 400 m long. It was covered with an uneven-aged spruce fir stand of about 20-m height. It was located near Fraser, Colorado, at an altitude of 3000 m. Vertical profiles of potential temperature deficit and wind speed at various locations on the slope showed approximate similarity when scaled by the height of the lowest inversion and by the average velocities and potential temperature deficits below that inversion. The latter varies in a nearly linear manner down the slope, and the inversion heights appear to be relatively constant over the range of net radiation rates observed. If a corresponding analytical similarity for the equations of motion and the sensible heat balance are assumed, the last two phenomena are found to imply that

$$U_m \propto \sqrt{\Delta\theta} \; ; \; \Delta\theta \propto R_o^{2/3} ; \; U_m / \sqrt{\gamma} \propto x \qquad (2.14)$$

where

- U_m is the average velocity
- $\Delta\theta$ the potential temperature drop down the slope
- γ the sine of the angle of the slope to the horizontal along the streamline
- x the downslope distance from the virtual origin of the flow measured along the streamline
- R_o the average net radiation loss on the slope.

Theoretical studies of katabatic winds have generally used four approaches. Humphreys (1920), Hann (1919), Kleinschmidt (1921), Jeffreys (1922), Prandtl (1942), and Defant (1949 and 1951) considered katabatic winds analytically. The analytical approach taken by Prandtl arrived at a relation for a steady-state wind, parallel to the slope, in terms of gravity, turbulent friction and turbulent conduction in a compressible atmosphere. If exchange coefficients were constant and the slope small, the equations of motion could be solved to obtain the velocity and temperature fields as functions of distance normal to the slope. His solutions were

$$\theta' (n) = \theta'_o \cos \frac{n}{\ell} \exp \left(-\frac{n}{\ell}\right) \tag{2.15}$$

$$n = \theta'_o \left(\frac{g\alpha K_m}{BK_h}\right)^{\frac{1}{2}} \sin \frac{n}{\ell} \exp \left(-\frac{n}{\ell}\right) \tag{2.16}$$

where θ'_o--temperature rise on the surface of the slope (n = o)

n--coordinate normal to the slope

K_m--eddy viscosity (assumed constant

K_h--eddy conductivity (assumed constant)

α--coefficient of thermal expansion

g--gravity

B--temperature lapse rate under undistrubed or motionless conditions

ℓ - $\left[4\left(\frac{K_m K_h}{g\alpha}\right) \frac{\operatorname{cosec}^2\phi}{B}\right]^{1/4}$

ϕ--slope angle.

Defant (1951) extended Prandtl's results by including time variations. Other researchers have extended Prandtl's approach by including height-dependent exchange coefficients or nonstationary conditions (Gutman 1953), and including the Coriolis force (Gutman and Malbakhov 1965, and Lykosov and Gutman 1972).

A numerical grid-point solution of the full primitive equations or a subset of them comprises the second approach (Thyer 1966). This technique has become more popular since the initial work of Thyer and, in time, may provide a reasonable estimate of the spatial distribution of drainage winds in different complex landforms (Leslie and Smith 1974 and Sklarew and Mirabella 1979).[a]

The third approach taken by Defant (1933)[b] and later Fleagle (1950), considered only the average flow within an identified cooled layer. All internal structure of the flow was eliminated and only variations in the time were considered. Fleagle presents two solutions: the first uses the dynamic theory including the inertia term; the second uses the equilibrium theory and excludes the inertia term. The first solution predicts that the mean velocity in the layer undergoing cooling

- is proportional to the net outgoing radiation
- is inversely proportional to the thickness of the layer undergoing cooling
- begins by varying periodically and gradually becomes constant
- is inversely proportional to the slope of the ground.

The last two predictions are due to the compressibility or adiabatic heating of the atmosphere and to the inertia of the air. The second solution shows that the drainage velocity is directly proportional to the net outgoing radiation, is inversely proportional to the thickness of the layer and slope, and approaches its final value more rapidly with a steep slope and large coefficient of friction than with a relatively flat slope and a small coefficient of friction. The rate at which equilibrium is attained depends upon the assumed form of the frictional force. A frictional force proportional to the square of the velocity gave fairly realistic results. Petkovsek and Hocevar (1971) extended Fleagle's ideas to include a stable background temperature stratification but in so doing obtained the anomalous result of the predicted katabatic velocity becoming infinite as the ambient stratification approaches adiabatic.

(a) See ASCOT--Information survey (1980) for additional material on numerical codes.
(b) Quoted by F. Defant 1949.

The fourth approach considers the theory of open-channel hydraulics (see Ball 1956, Ellison and Turner 1959, and Manins and Sawford 1979). Manins and Sawford used a model that replaced the detailed vertical structure of the flow by a quiescent stably stratified environment and an equivalent flowing layer subject to sustained cooling, surface stress, and interfacial entrainment. Their results showed that interfacial entrainment is the dominating retardation mechanism of the flow and that surface stress may be relatively unimportant. Steady solutions show that katabatic winds are essentially super critical on all practical slopes (slope angles > 0.1°) and are affected by ambient stratification only at large distances. This latter method has been utilized for predicting drainage flow down a slope from a cold source, such as the surface winds from Greenland, the Antarctic littoral, and glaciers (Ball 1957 and 1960; Gutman and Malbakler 1965; Lettau 1966; Streten 1963; and Streten, Ishikawa, and Wendler 1974.)

2.4.3 Mountain and Valley Winds

The wind changes daily along the axis of large valleys. In the day, from about 0900 to 1000 until sunset, an upvalley, or valley wind blows. At night, an opposite downvalley or mountain wind appears that continues into the early morning hours after sunrise. These winds occur most frequently during persistent high pressure situations in summer and are thus a typical fair-weather phenomenon. Mountain and valley winds can occur on cloudy days and in winter when they modify the general wind. Mountain and valley winds are best developed in wide and deep valleys, such as in the Alps, but they also occur in small canyons. The shape of the valley's cross-section and the inclination of the valley bottom are of little influence on these winds.

2.4.3.1 Observations

Numerous observations and investigations of the mountain-valley wind system are discussed, although no attempt has been made to include all studies. [See Buettner and Thyer (1966) and Orgill (1968) for literature surveys.] Mountain and valley winds were observed and described as early as 1840, with some of the first extensive observations in the Alps reported in the early

1920s and 1930s.[a] Ekhart (1931, 1934, and 1936) conducted aerological investigations in W- and V-shaped valleys in the Innsbruck region of Austria. During the 1940s and 1950s Hewson and Gill (1944) conducted field experiments at Trail, British Columbia, Canada, in the Columbia River valley. Rather detailed observations on the diurnal variation of the slope-valley winds as well as the cross-valley winds were made at different locations in the valley. Cross (1950) later supplemented the results of the study and discovered that the usual diurnal wind pattern prevailed in the valley especially during the summer.

During the 1960s, several field studies took place in the United States, Japan, Norway, Canada, United Soviet Socialist Republic, and South Africa. Buettner and Thyer (1962 and 1966) investigated the mountain-valley and slope winds near Mt. Rainier, Washington, for four summers. A valley wind and a mountain wind occurred rather regularly under relatively light geostrophic winds. Both wind systems were compensated by a return flow or anti-wind at a higher level (see Section 2.4.3.3 for anti-winds). The layers occupied by the two flows were of approximately equal thickness, and the boundary between them was generally at or below ridge height. Horizontal wind speed in these two layers was greatest slightly below the center of each layer. Speeds reached a maximum in early afternoon and just before sunrise. Vertical transport of air between the two layers appeared to be localized in the neighborhood of the ridges (see Figure 2.21). Usually when a well-developed wind system occurred in one valley, well-developed systems tended to occur in other valleys in the same area. Buettner and Thyer also suggested that a well-developed valley-slope wind systems would require that:

- the valley fully develop radiation-caused temperature differences,
- the temperature gradients cause the respective winds, guided only by the valley walls
- adjacent valleys not interfere

(a) For an excellent survey of the early work see Hawkes (1947).

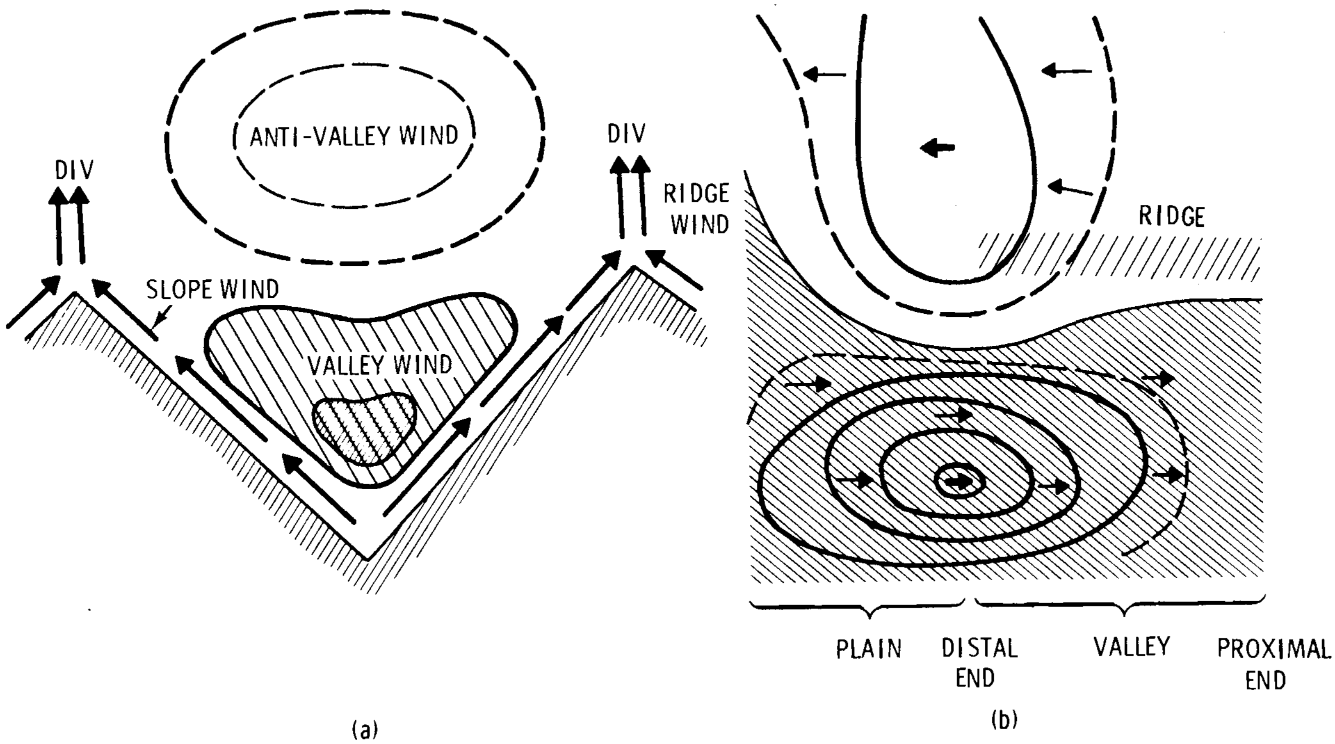


FIGURE 2.21. Model of Daytime Valley Wind Based on Field Measurements Near Mt. Rainier, Washington. Nighttime mountain wind and anti-winds are essentially reversed from those shown (adapted from Buettner and Thyer 1966).

- valley be continuous, long, deep, steep, and of uniform surface configuration
- the valley's proximal end be a plateau or a mountain, and not a pass, and the distal end a plain.

Extensive field measurements on the mountain winds were conducted in seven valleys in southwestern Vermont (Davidson and Rao 1963). The existence of the daytime upvalley and upslope winds were not confirmed by observations because of the existing meteorological conditions and particular valleys. The nighttime valley-plain wind dominated the wind pattern in a valley under clear weather conditions when the upper level (geostropic) flow was less than 5 to 10 ms^{-1}. The downslope wind occurred even under extreme wind speeds (>10 ms^{-1}) at the ridge line. In short valleys with well-defined heads and slopes the valley wind primarily drained the slope wind system. In longer valleys with poorly defined heads and divides, the temporal variations of the valley-plain wind system were caused by a large-scale pressure gradient directed from valley to plain.

The vertical extent of the valley-plain wind was finite and a function of the height of the ridge line and speed of the prevailing flow. The height of the wind system (H) for three valleys was approximated by

$$H = H_o \ (1 - 0.5u) \tag{2.17}$$

where H_o is the height of the ridge line and u (ms^{-1}) is windspeed at maximum ridge-line height. Generally, the maximum speed was reached within the lower half of the layer. The height of the maximum speed (h) averaged at about one-half of the wind system or h/H = 0.50. The range varied from 0.2 to 0.8. Figure 2.22 shows some typical windspeed profiles for the valley-plain wind.

Considerable downvalley acceleration of the wind may exist with the valley-plain wind system. In the case of the Vermont valleys, the average downvalley velocity was

$$\bar{u} \propto \sqrt{x} \tag{2.18}$$

where x is the down-valley distance up to 15 km from the head of the valley. The area of maximum wind descended almost to the ground as the air issued from the valleys.[a]

Sterten and Knudsen (1961) and Sterten (1963) described two extensive field programs on mountain-valley winds in Norway. Their results suggest a close relationship between the mountain-valley wind system and interdiurnal variations in atmospheric pressure. Sterten theorizes that the mountain wind causes a pressure wave to move down the valley during the night. After

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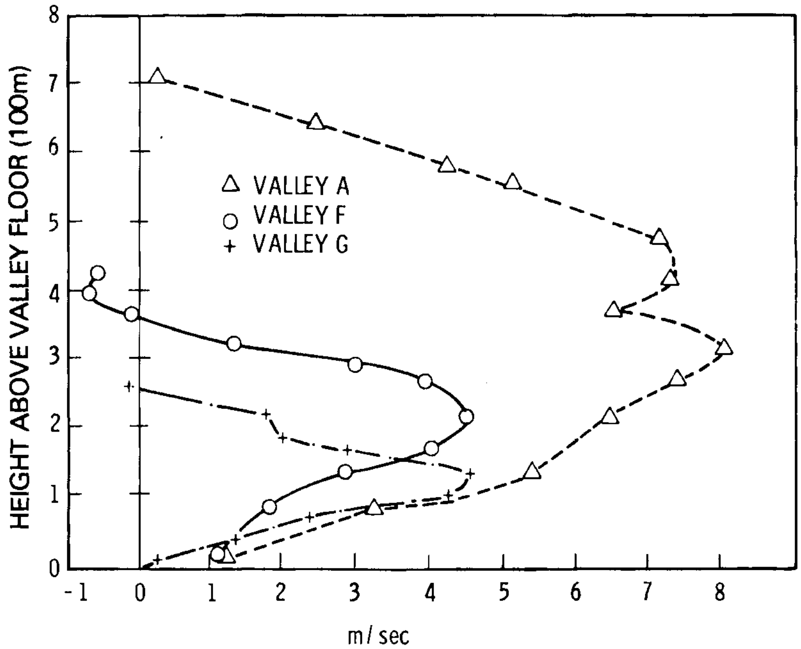
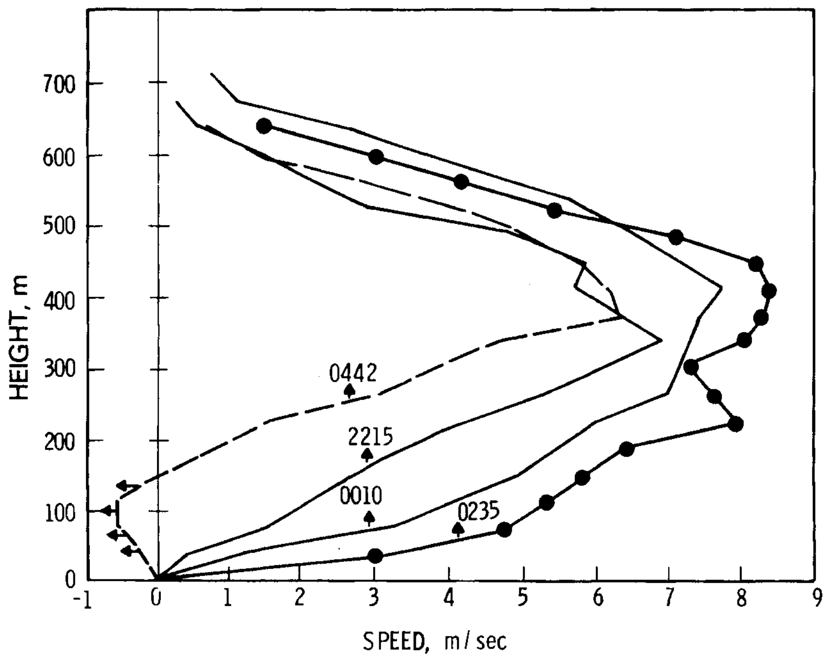
(a) Similar features were observed by Hales (1933) in canyons of the Wasatch Mountains in Utah.



(a)



(b)

FIGURE 2.22. Velocity Profiles During the Mountain or Valley-Plain Wind: a) Profiles Observed in Different Vermont Valleys, July 1959 and b) Time Variations of the Valley-Plain Wind Profile as Observed in Valley A, July 1959. Each curve is the average of two or three consecutive observations centered at the indicated hour (adapted from Davidson and Rao 1963)

sunrise, the mountain wind suddenly breaks up and partly disappears in the ground layers although it exists at higher levels of the valley. The transport of cold air causes a gradual drop in pressure in the upper valley region, but in the lower region the pressure rises. These changes in the pressure gradient after sunrise are suggested as a factor for the development of the valley wind. Tyson (1969) reported similar observations in South Africa.

Petrosyants and Chanysheva (1963) used summer observations from 15 valleys in the United Soviet Socialist Republic (U.S.S.R.) to discuss the relationship of valley circulation to the dimensions of the valley. In small valleys, 15 to 20 km in length, the valley wind begins around two hours after sunrise. In valleys of several kilometers in length, the wind changes direction 5 to 6 hr after sunrise. Generally, the highest speed of the surface valley wind was observed when the valley length (L) and width (W) ratio was $12 \leq L/W \leq 18$. The ratio for maximum valley windspeeds was $7 \leq L/W \leq 10$. The ratio of the mountain (downvalley) windspeed (V_m) to the valley windspeed (V_v) was found to be independent of the valley direction and less than unity ($V_m/V_v < 1$) for closed valleys; that is, the valley wind was stronger. In open valleys the ratio varied according to the direction of the valley. In valleys, opening to the southwest, west and northwest the ratio V_m/V_v was less than 1 so the valley wind prevailed; in valleys opening to the north and northeast V_m/V_v was greater than or equal to 1; i.e., mountain winds prevailed.

Thompson (1967) made a micrometeorological investigation of the formation, maintenance and dissipation of ground layer inversions and related thermal circulations near the mouth of a small canyon in the Wasatch Range of northern Utah. One of the main conclusions of the study was that meteorological events are similar for small and large valleys but occur earlier in the day and undergo a more rapid development in the smaller canyons.

MacHattie (1968 and 1970) studied the local winds and temperatures in a valley oriented north-south in Canada. Prominent diurnal cycles of the cross-valley component were observed. The component of wind along the main valley had a more complex pattern because of gradient wind interference in the afternoon when convective activity was greatest.

A number of field efforts were conducted in valleys during the 1970s to evaluate existing or potential air quality and pollution problems. A rather extensive field survey was made on local winds in the industrialized Kanawha Valley in West Virginia (USNAPCA 1970). The study examined the effect of local industrial heat sources and the river on the local valley wind system (Figure 2.23).[a] A cooperative field experiment (Fox et al. 1976) was conducted in December 1975 in a highly developed mountain valley in the Central Rocky Mountains. Classical mountain-valley wind systems were observed to occur a large portion of the time. The angle of the upper-level wind to the valley axis was found to be critical in determining the degree of interaction between ground-generated and larger scale circulation.

Kao, Bidwell, and Lee (1974) investigated the topographical effect on air motion in the Salt Lake valley. This open valley has a unique feature of

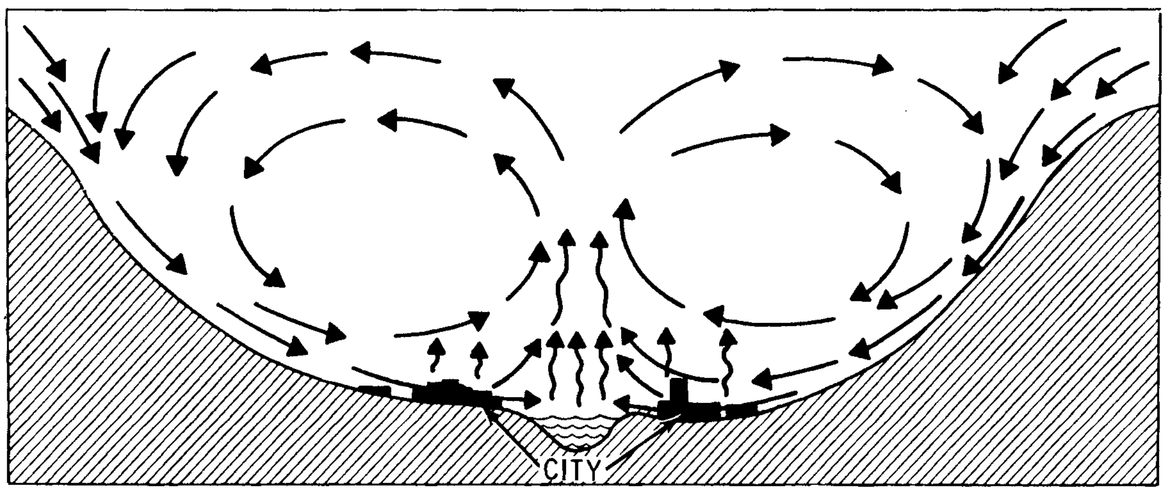


FIGURE 2.23. Idealized Cross-Section of Nocturnal Slope-Valley Winds and Their Modification Caused by a Warm River and Local Industrial Heat Sources

(a) The effect of cold air drainage on Calgary's (Canada) urban heat island has been reported by Nkemdirim (1980).

having mountain-valley, lake breeze, and industrial heat island circulations present in proximity to one another. The mean motion in the valley is strongly affected by the mountain-valley winds and shows a southeasterly flow in the evening and early morning, a northwesterly flow in the afternoon and a transitional flow in the late morning and after sunset. The mountain winds were generally associated with a horizontally divergent flow.

Ryan and Brown (1978) studied the local winds in a north-south canyon in the San Bernardino Mountains of southern California using a network of automatic telemetered stations. Observation stations were placed on slopes, in canyons and on ridges at different elevations and aspects so that the effects of topography on weather variables could be isolated from mesoscale and synoptic scale weather influences. Analysis of daytime and nighttime winds showed that winds tend to switch directions from day to night. Often the reversal was not 180°, but depended on the canyon orientation both below and above a particular station. Morgan and Slusser (1978), and Carroll and Baskett (1979) present information on mountain-valley and slope wind systems in the Sierra Nevada. Ruff (1979), Steffen et al. (1979), and Knuth and Jensen (1978) have presented some preliminary data on the local wind systems in the Geysers Geothermal Resource Area north of Santa Rosa, California.

Given the results of these various field observations a statement from Buettner and Thyer (1966) seems fitting.

> There is neither a perfect valley nor a perfect gradient wind-free weather condition; there will never be enough synoptic three--dimensional observations in order to show all parts of a valley wind system clearly. The best we can do is to establish a model of happenings under ideal conditions. This model should be as close as possible to experimental facts and to theory (p 143).

2.4.3.2 Theories

Thyer's (1966) review of several older theories on mountain-valley winds indicated that the temporal and spacial variations of three-dimensional mountain and valley systems were not well explained. To quote Thyer:

An ideal theory of valley and mountain winds would give, as functions of all three space coordinates and of time, all three velocity components, the temperature, pressure, and density, in the valley, above it and over the adjacent plain. Slope winds, valley/mountain winds, their respective anti-winds and vertical currents would all be explained simultaneously, and this would all be deduced from the minimum initial conditions of the geographical location, the

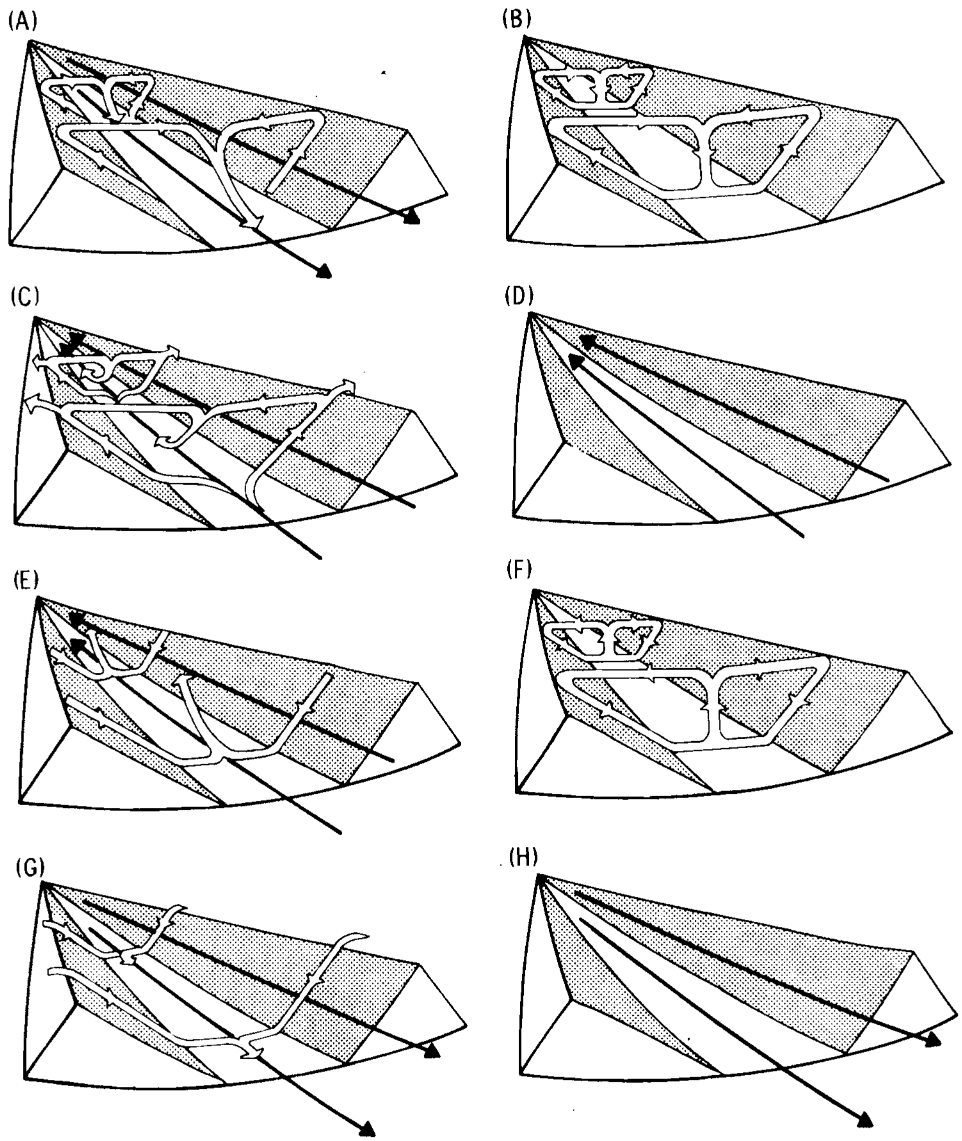


FIGURE 2.24. Schematic Illustration of Idealized Diurnal Variations of the Air currents in a Valley According to Wagner--Defant Theory: a) Sunrise, b) Forenoon, c) Noon and Early Afternoon, d) Late Afternoon, 3) Evening, f) Early Night, g) Middle of Night, and h) Late Night to Morning (adapted from Defant 1951)

> topography of the area, the nature of the surface of the ground (especially its albedo, specific heat and thermal conductivity), the insolation on the surface, or radiation from it, as a function of position and time, and the prevailing atmospheric conditions, such as gradient wind, stability, humidity, and cloud. Further it would not ignore the fact that actual valleys are finite in size (p. 321).

Prior to Thyer's work the principal theory for the mountain-valley and slope wind systems was the Wagner-Defant qualitative theory (Figure 2.24). This theory showed that two wind systems were required to preserve the stationary condition in the valley. One of them was the horizontal inflow (outflow) of the valley (mountain) wind; the other the thermal slope wind system that takes care of the outflow (inflow) over the flanking ridges. Observations suggest that slope winds existing near the walls of the valley are the principal means of vertical transport of air and are therefore an integral part of the wind system. As a result of the principle of continuity, valley or mountain winds are then, in part, a consequence of such slope winds. Slope winds are essentially the result of a baroclinic condition, caused by horizontal temperature gradients. These temperature gradients exist when unequal heating of air occurs at different points at a given altitude.

Davidson and Rao (1958) and Buettner and Thyer (1966) have noted differences between observations and the Wagner-Defant theory. Among these were:

- Some valleys may not have either the daytime slope or valley wind (Davidson and Rao 1958).
- No periods of slope winds only, or valley winds only occur at morning and evening, only irregularities occur in the turn-over period caused by the structure of real, not ideal valleys (Buettner and Thyer 1966).
- No measurable vertical wind components occur in mid-valley at day or night (Buettner and Thyer 1966).
- The slope winds at daytime that end in strong vertical currents at the ridges feed the anti-winds (Buettner and Thyer 1966).

Thyer (1966) developed a simplified mathematical model from the primitive equations for simulating valley-slope winds in a idealized V-shaped valley. Characteristics reproduced by the model included a thin layer of slope wind, updrafts over the ridges, valley or mountain wind below ridge level having a maximum speed over the trough and an anti-wind layer. Figure 2.25 is a comparison of Thyer's numerical results to field observations for a valley wind in the Carbon River valley. Asymmetrical aspects of the valley wind could not be simulated, and the gradient wind and Coriolis force were neglected.

Thyer and Buettner (1961) also attempted to simulate valley-slope winds by laboratory experiments for an ideal V-shaped valley. Simulation of day conditions (heating) and night conditions (cooling) to the slopes forming the valley was obtained by using heating tape and solid carbon dioxide.

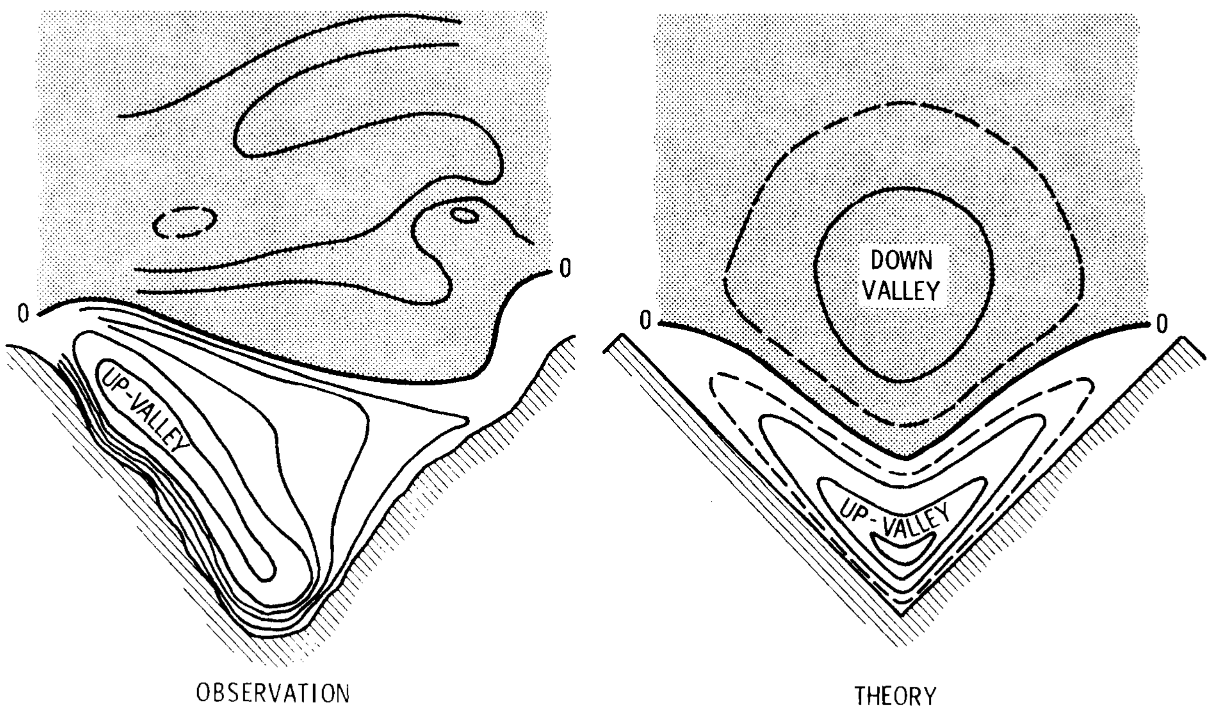


FIGURE 2.25. Comparison of Field Observations (1300 PST, July 9, 1959) of Longitudinal Wind Component in the Carbon River Valley with Theoretical Results from Thyer's Numerical Model (adapted from Thyer and Buettner 1966)

Similitude requirements were ignored. Their conclusions indicated that a sloping valley floor is not necessary for a valley-slope wind system, but the slope winds are of importance.

Several models of thermal and terrain-induced perturbation of near-ground flow have been developed recently. A few of these are Lantz (1972); Fosberg, Marlatt, and Krupuak (1976); Ryan (1977), Sklarew, Wilson, and Fabrick (1977); and Yamada (1978 and 1979).[a] Although the results from these models look encouraging the problem of mountain-valley winds still remains to be solved and validated.

2.4.3.3 Anti-Winds or Compensation Wind

Immediately above the valley or mountain wind, a counter-flow is sometimes observed that is in approximately the opposite direction. Buettner and Thyer (1966) called this the anti-wind or compensation wind. The anti-wind, when it exists, usually has a definite upper limit, and above this limit is the gradient wind. Most of the information on anti-winds is from Buettner and Thyer's observations near Mt. Rainier, Washington.

Anti-winds may not exist for all valleys. When they occur above ridge level, they may deviate considerably from the lower valley or mountain wind because of the influence of gradient winds, bends in the valley across which they can take shortcuts, and mixing with anti-winds of other valleys that are not parallel. Anti-winds generally occupy a region of about the same vertical thickness as the lower valley-mountain wind. Sometimes in the presence of strong gradient winds the thickness of the anti-wind layers is reduced to below ridge level. Occasionally, the anti-mountain wind is depressed wholly below ridge level; however, this effect is more likely to occur at night than in the day.

The transition between anti-valley and anti-mountain (upvalley) winds is not always simultaneous with the corresponding lower valley mountain wind transitions. Generally the former tends to occur somewhat later, with a delay of a few minutes to some hours. A poorly developed anti-wind may both start

(a) See ASCOT Information Survey for additional modeling efforts.

later and finish earlier than the corresponding lower valley-mountain wind. If the anti-mountain wind is depressed, it may merge imperceptibly into the valley wind of the following day. Valley winds and their anti-winds, if not modified by gradient winds, are somewhat faster than the mountain wind system. Otherwise both systems tend to mirror each other as shown in Figure 2.21.

2.4.3.4 Surges

The nocturnal mountain winds show a considerable downvalley acceleration of the wind as it proceeds from the head of the valley to the mouth of the valley. Superimposed on this downvalley acceleration of the wind are quasi-periodic fluctuations in the maximum wind called "surges." Surges in the mountain wind have been reported by Pollak (1924), Heywood (1933), Buettner and Thyer (1966) and Tyson (1968). Fleagle (1950) predicted a periodically varying drainage current in his dynamic theory of air drainage. Common periods of surges have varied from 20 min to 75 min.

According to Fleagle and Tyson, surging in the mountain wind is due to compressibility of the atmosphere and to the inertia of the air. The pressure gradient producing the downvalley wind is weakened by adiabatic heating and local horizontal divergence in air accelerating down the valley in a surge. The resulting deceleration of the mountain wind occurs until radiational cooling again increases the pressure gradient and the cycle repeats itself. Apparently, similar periodic fluctuations in the valley wind (daytime) are not as noticeable, are absent, or have not been reported.

2.4.3.5 Wind Profile Instabilities

Davidson and Rao (1963) discovered a major small-scale unrest in the mountain-wind profile in the layers below the level of the maximum wind, generally below 200 m. Figure 2.22b shows a rather dramatic breakdown or instability in the lower portion of the wind profile. Reversed airflow and cross-valley wind components greater than the mountain wind component appear to be characteristic of this type of instability. The development of large wind shear (0.02/sec to 0.06/sec) near the ground was tentatively suggested as a possible mechanism for initiating this type of instability but the actual mechanism for this phenomenon has not been positively identified.

2.4.3.6 Transitional Periods

In early morning or late afternoon, the vertical structure of airflow within any valley may be complicated. In the case of the early morning nocturnal mountain winds, disipation may proceed from either above (i.e., from ridge level downwards) or from below (i.e., ground upwards).

Pollak (1924) first observed the dissipation of the mountain wind from above in a valley near Trento, Italy. His observations indicated that the height of the mountain wind decreased about 700 m between sunrise and 0900 or 1000. Hewson and Longley (1944) indicated that the top of the mountain wind steadily lowered at a rate of 70 ± 20 m hr^{-1} in the Columbia River Valley in British Columbia. Davidson and Rao (1958, 1963) observed a decay rate of 80 to 120 m hr^{-1} in valleys in Vermont. Ayer (1961) estimated that this decay rate was 40 to 50 m hr^{-1} in the Carbon River valley of Mt. Rainier, Washington. However, Buettner and Thyer (1966) found cases when the transition period would take place at all levels within a few minutes, but slightly earlier at the surface than at higher levels. Tyson (1969) indicated for some individual valleys in South Africa that the valley wind often initiated at or near the entrances of the valley and advanced upvalley undercutting the decaying downvalley mountain wind.

Whiteman and McKee (1977, 1978, 1979, and 1980b) studied the evolution and interrelationship of wind and temperature structures during clear but light wind periods. In the Gore River Valley and six other mountain valleys of western Colorado, the nocturnal temperature inversion descended into the valley at a rate of about 120 m hr^{-1} after sunrise. The windfield was characterized by upvalley winds in a nearly dry adiabatic layer above the inversion, light downvalley winds in the cold air below the inversion, up-valley winds at the surface and up-slope winds on the sidewalls (Figure 2.26). During late afternoon and early evening, the nocturnal ground-based temperature inversion developed rapidly to a depth of 175 m in less than two hours. The inversion continued to deepen and was accompanied by a strong cooling of the air in the total depth of the sounding. Winds were very light in the valley during the night.

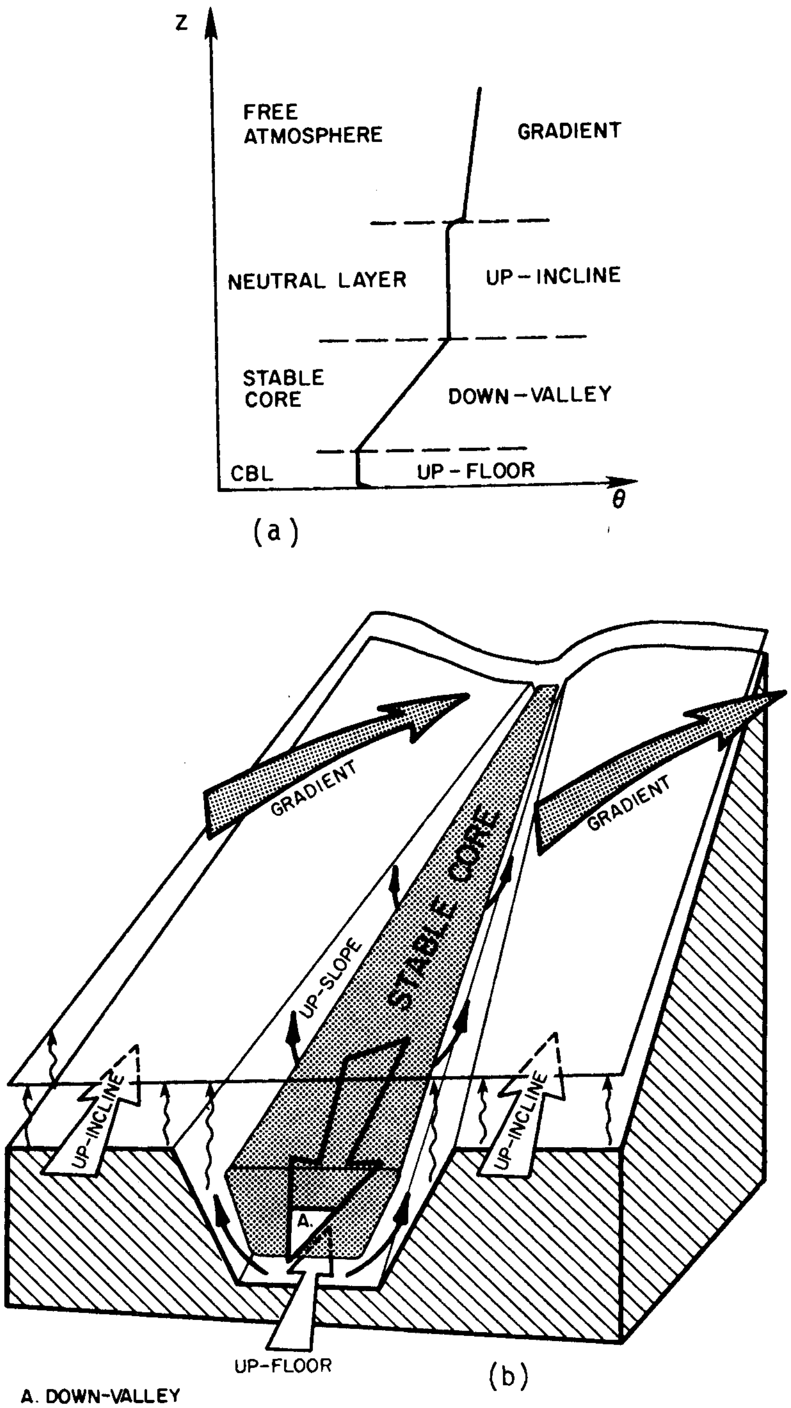


FIGURE 2.26. Temperature and Wind Structure during Valley Inversion Dissipation: a) Correspondence Between Temperature Layers and Wind Fields and b) Observed Wind Systems at Mid-Morning During Inversion Dissipation (courtesy of C. D. Whiteman)

At least three hypotheses have been invoked to explain the dissipation of nocturnal drainage winds. Ayer (1961) suggests that the lowering of the top of the drainage current is due to convective mixing established by morning insolation, which is faster on the upper slopes of a valley than at the bottom, combined with the removal of the source of fresh cold air by the drainage of cold (stable) air out the end of the valley. Davidson and Rao (1958) postulate that instability develops at ridgetop level and the general wind responds quickly and descends into the valley. The more unstable the air at the ridgeline, the deeper the penetration of prevailing flow into the valley. Shieh (1971) used a simple one-dimensional model to simulate the dissipation of the drainage current from above and apparently confirmed some aspects of Davidson and Rao's hypothesis. Whiteman and McKee (1978, 1979, 1980) indicate that the descent into the valley of the stable layer associated with the drainage layer and its eventual dissipation are the result of convection forming along the slopes because of solar insolation that carries the entrained air up the slopes and out from under the cold air layer. As air is removed by the upslope wind system the inversion top descends and the upvalley winds progress deeper into the valley until the inversion and drainage current is completly dissipated. All three hypotheses may be valid depending upon various factors such as the valley and time of year.

Thompson (1967) conducted a study on the temperature and wind distribution during transition periods in and near the mouth of Red Butte Canyon in the Wasatch Range of northern Utah. Data were taken during the summer of 1957 and showed that a basic pattern occurred on 90% of the days. The main features of the temperature and wind behavior patterns were:

- formation of a thin film of cold air on the floor of the canyon and near the canyon slopes
- rapid wind shift from up-canyon to down-canyon wind with variable wind direction before the establishment of the persistent down-canyon wind
- extremely rapid cooling within the canyon during the first hour of down-canyon wind with slight cooling after the first hour

- rapidly increasing windspeed accompanying the cooling, followed by steady or slightly decreasing wind the remainder of the night
- presence of a core of cold air in the canyon as morning heating progressed
- rapid wind shift from down-canyon to up-canyon in the morning.

Some of these characteristics are depicted in Figure 2.27. Thompson indicates that the important difference in events between large valleys and small canyons is the earlier and more rapid development in the smaller canyons.

2.4.4 Interaction Between Wind Systems

The prevailing geostrophic wind patterns and other local wind systems strongly influence local winds in a particular coastal (lake) or rough terrain region. If the interaction between local and other airstreams is weak in complex landforms, the individual wind systems are essentially decoupled from each other and the airstreams exist as if the neighboring airstream is practically not present However, if the coupling is strong, one airstream may dominate the other, or in some cases the amalgamation of the different airstreams may create an airstream of different properties or characteristics. Two types of interactions involve synoptic and local airstreams.

2.4.4.1 Synoptic and Local Airstream Interaction

The development of a land-sea breeze (LSB) or land-lake breeze (LB) is very dependent on the strength of the prevailing synoptic airflow. If the prevailing synoptic flow is too strong then the land-sea breeze will not develop; however, a whole spectrum of modified circulation patterns can develop if the prevailing airflow is below or near the critical cutoff wind velocity.

Leopold (1949), Estoque (1962), Lavoie (1967), Mendonca (1969), and Neumann (1977) examined the interaction of the land-sea breeze of a coast or island and the prevailing airflow. Estoque developed a mathematical model based on the hydrostatic equations of motion to evaluate different geostropically balanced initial conditions on the development of the sea-breeze circulation. Neumann showed that the rate at which the direction of the land-sea breeze turned resulted from the Coriolis parameter, a horizontal mesoscale

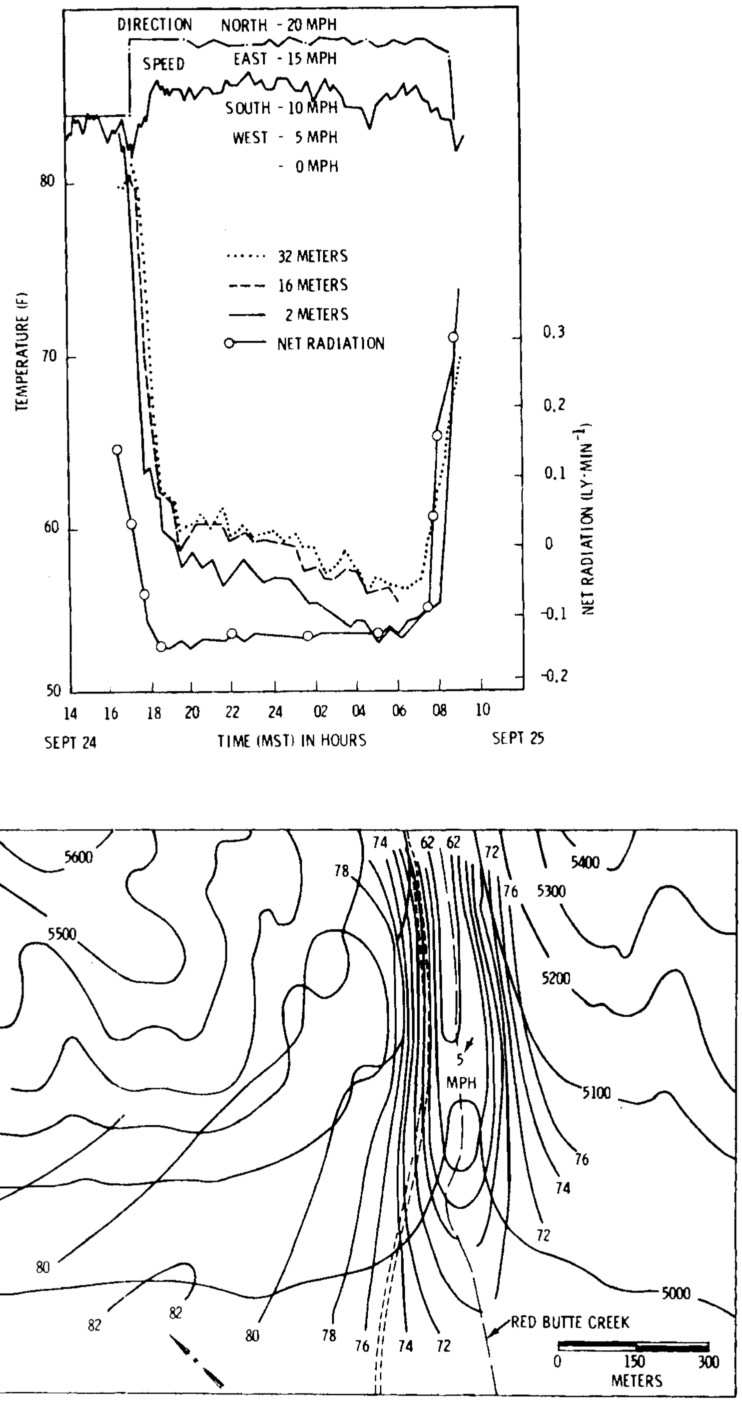


FIGURE 2.27. Drainage Winds Out of a small Utah Canyon: a) Variation of Winds, Temperature and Net Radiation at the Mouth of Red Butte Canyon--Time: September 24-25, 1957. b) Temperature distribution at 2 m above the surface at the mouth of Red Butte Canyon--Time: 1740 MST, September 24, 1957 (adapted from Thompson 1967)

pressure gradient (approximately equivalent to the diurnal heating and cooling of the land relative to the sea) and the horizontal large-scale pressure gradient. The last two items vary in magnitude and sign and essentially modulate the rate of turning of the land-sea breezes at a coast. Figure 2.28 shows a few typical interactions between land-sea breeze phenomena and prevailing airflow for different islands in the Hawaiian chain (Leopold 1949).

The prevailing (synoptic) airstream may interact with the local winds of a valley in three general ways: when winds are light, moderate to strong, or at an angle to the valley. During periods when the prevailing or geostrophic wind is light or negligible, only minor modifications to the local winds result. If other conditions are favorable, strong local wind systems could develop.

A second way occurs during periods in which the prevailing wind is moderate to strong and the direction is approximately parallel to the valley axis. Probably in most situations the local wind system would be overpowered by the prevailing flow or at least strongly modified by it. However, if local winds are about the same magnitude as the prevailing airflow, different interactions could occur. In Figure 2.29, a pronounced elevated shear zone is formed by the interaction between the channeled synoptic wind (plus valley wind component) and a katabatic wind that flows off the slope of the Continental Divide. The slope of the shear zone or microfront depends on the relative densities of the two opposing air currents (Wilkins 1956; Cate 1977).

A third way occurs when the prevailing wind is moderate to strong and the direction is at an angle to the valley or ridge axis. In this case, the valley width-height ratio and atmospheric stability affect the interaction. Local wind systems formed in deep narrow canyons and valleys will be less affected by the prevailing airflow. This would be particularly true for the nocturnal drainage wind that could be essentially decoupled from the higher prevailing airstream. Local wind systems in wide valleys could experience greater interactions with the prevailing or synoptic airstream (Countryman and Colson 1958).

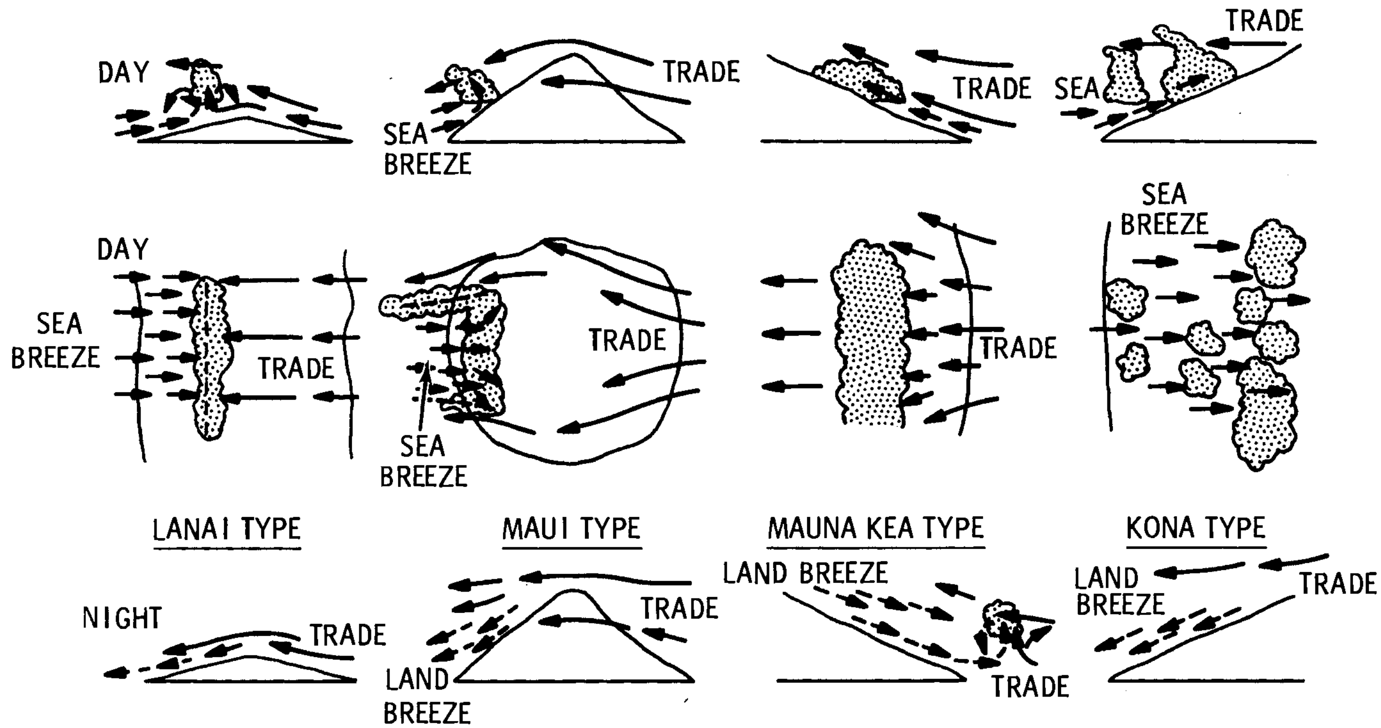


FIGURE 2.28. Typical Interactions Between Island Land-Sea Breezes and the Large-Scale Prevailing Airflow (adapted from Leopold 1949)

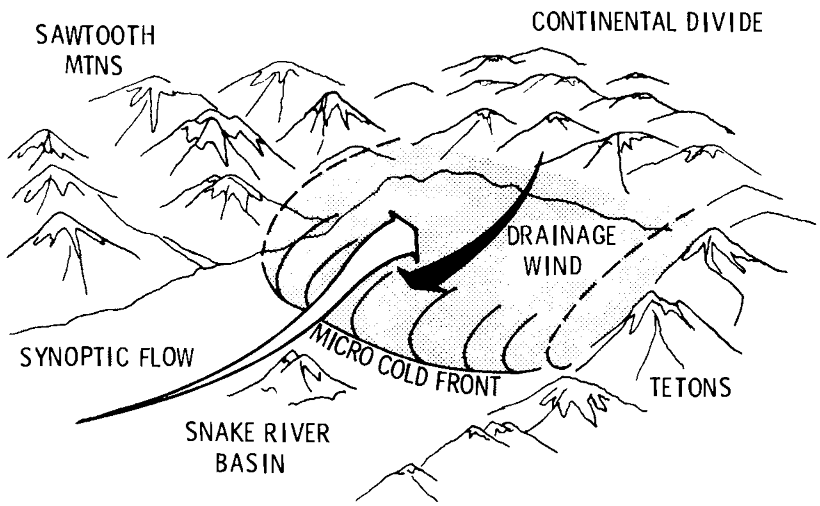


FIGURE 2.29. Schematic Illustration for the Snake River Basin (Idaho) Indicating the Interaction of a Local-Microfront (drainage wind) with the Valley Synoptic Flow

Interaction between synoptic or prevailing airflow and local upslope winds occurs during the summer in east-facing canyons and slopes of the coastal mountains of California and Oregon, and the Cascade mountains of Washington. During the early afternoon, the daytime upslope or canyon winds on the east side of the mountains are frequently replaced by a moderately strong downslope wind that can, in some locations, extend several kilometers inland. The mechanism for the development of such winds is not fully understood but Schroeder (1961) suggests that the winds are in part due to the marine air spilling over the mountain crests in response to a trough aloft and the seasonal monsoon-sea breeze and west-to-east pressure gradient. High elevated temperature inversions and some degree of instability in the lower atmosphere also contribute to this condition. Staley (1959) also cites the inland progression of the large-scale sea breeze as a reason for the change in wind direction. Fosberg and Schroeder (1964) and Fosberg (1969) suggest that the combination of instability and lee wave development may be the mechanism for causing the downslope winds in the afternoon. The combination of leeside separation of the airflow and the development of lee waves may also be mechanism for such winds (see Section 2.22).

Figure 2.30 shows an ideal sequence of events depicting the possible development of the afternoon downslope winds. With the sun warming the east slopes of the mountains the upslope wind system develops (Figure 2.30a,b) and eventually separation occurs on the lee side forming a leeside eddy. Later in the morning, the west-to-east pressure gradient increases, valley winds develop on the west side and the eddy retreats or moves off to the east allowing a wave to develop just east of the mountain crest (Figure 2.30c). Later in the early afternoon this combination of factors results in the downslope wind moving inland (Figure 2.30d). In the evening if the marine air is still flowing, it may merge with the nighttime drainage flow from the mountains (Figure 2.30e). Since the air is normally quite dry on the east side, the marine air is quickly modified by adiabatic and radiational heating. Very few if any clouds are associated with the development of these afternoon downslope winds. A similar sequence of events may occur in the summertime in the Rocky Mountain chain but cloud and thunderstorm development over the mountains affects the nature and timing of the downslope winds (Dirks 1969).[a]

An analytical model was developed by Tang (1976) to investigate the interaction between the large-scale prevailing airflow and the thermally induced slope wind circulation in an idealized valley. With the prevailing airflow perpendicular to the valley a separated circulation cell is formed over the lee slope under daytime unstable conditions. On the windward slope the prevailing wind persists. For the nighttime case, a weak elongated cell is developed over the windward slope. This is attributed to the development of the drainage wind over the windward slope, which counteracts the relatively weak prevailing wind above ground. On the lee slope the drainage wind tends to counteract a frictionally driven separated cell leading to a weakening of this cell (Figure 2.31).

2.4.4.2 Interaction Between Local Flows

Local airstreams generated in neighboring valleys may interact. This interaction may result in wind direction and speed changes in the airstreams, in delay or acceleration of the onset of local winds, and replacement of one

(a) Also see Figure 2.8 and related discussion.

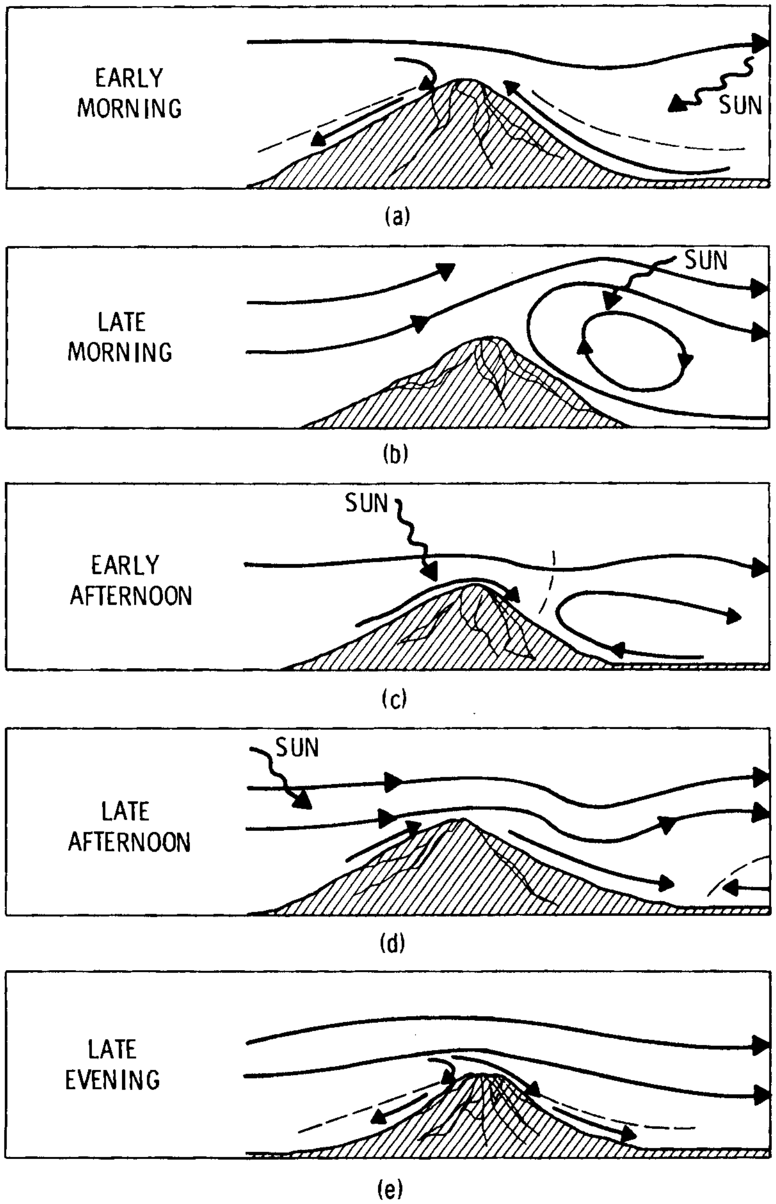


FIGURE 2.30. Sequence of Possible Events Depicting the Development of Downslope Afternoon Winds East of Mountain Ridges During Summertime--No Cloud Development

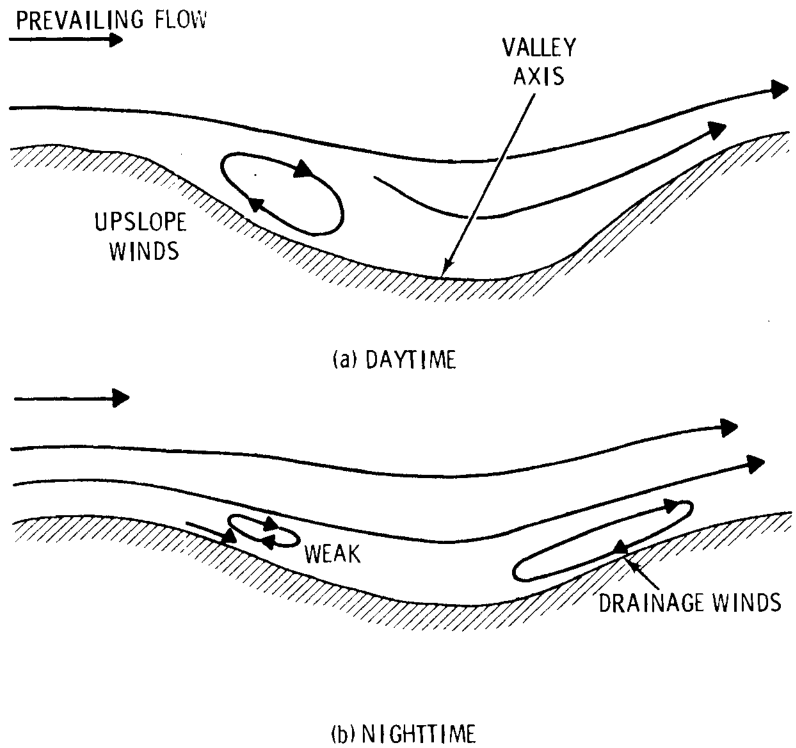


FIGURE 2.31. Interaction of Prevailing Airflow, and Thermally Induced Slope Wind Circulation According to Tang's (1976) Analytical Model

local airstream with another. Few systematic observations of interaction between local airstreams have been conducted, except for the Maloja wind.

The Maloja wind (Defant 1951 and Huschke 1959) named after the windshed between Engadine and Bergell, Switzerland, is a down-valley wind resulting from the valley wind of one valley reaching over a pass into another valley. This wind appears particularly well developed. The development of this anomaly is probably determined in part by which one of the valleys involved has the larger diurnal temperature amplitude and thus effectively extends its circulation into the other valley across the pass. Maloja-like winds have been reported in the United States by Buettner and Thyer (1966) and DeMarrais et al. (1968).

At a confluence of two or three valleys, the valley wind of one might draw in that of another and cause a phase shift in the occurrence of the maximum valley wind speed (Buettner and Thyer 1966). The presence of tributary canyons within a main valley may lead to aspirated airstreams caused by a Bernoulli effect between the main flow in the valley and the airstreams from smaller canyons. Confluence of side valleys may lead to a merging of two individual local wind systems of different thermal and wind characteristics (Gedayloo et al. 1980).

If the head of a valley is terminated by a glacier, an interaction between the local downvalley glacier wind and the upvalley wind can occur during the daytime (Geiger 1966). Ridgelines are the terminators of upslope flow from valleys. These air currents continue as thermals and eventually cause convective clouds. Clouds will cast shadows on slopes and in valleys and will release quantities of precipitation-cooled air that may form temporary local air currents that can interact with the valley-slope wind system. Local land-lake breezes that may form along large lakes in complex landforms could interact with local valley-slope winds. Observations of this type of interaction are apparently scarce (see Section 3.2.2.1).

2.4.5 Summary

In general, local winds have been widely studied and observed but the data base has not been organized and is variable in content because of the different characteristics of the sites and studies. The physics of local winds is reasonably well understood but present and future analytical, numerical and physical models of local winds will require careful comparison with observation before their validity can be established.

Two aspects of local winds that are in need of considerably more research are transitional periods and interaction between synoptic and local winds. Studies are needed on the physical processes, modeling and the collection of field data.

The understanding of local flows depends greatly on information on the heat balance at the surface and the radiative and turbulent transfer of the surface heat to the boundary layers. Studies on these physical processes in complex landforms is seriously lacking and should be remedied in the future.

2.5 STAGNATION AND NATURAL VENTILATION

Stagnation is an atmospheric event in which the following characteristics are usually observed: 1) persistence of a quasi-stationary anticyclonic weather pattern, 2) no significant precipitation, 3) low mixing depths (1500 m or less) and 4) mixing layer average wind speeds of 4 ms^{-1} or less. The above conditions must continue for at least two days (Holzworth 1962, 1964, 1971, 1972 and Hosler 1961). Holzworth evaluated stagnation events over the United States. His results indicated that in the eastern portions of the country, autumn is the season in which stagnation most often occurs. In the West, the most common season is winter, except in the northern part where it is autumn due to the frequency of winter storms. As a result, the meteorological potential for air pollution is clearly much greater in the west than in the middle or eastern parts of the United States (Figure 2.32). In the mountainous West and East, valleys and basins, particularly those with steep slopes, favor stagnation because of the confining nature of their surrounding ridges.

2.5.1 Valleys

Olsson and Peterson (1971) illustrated the nature of the stagnation and natural ventilation problem in a valley using the Columbia-Willamette River valley in Oregon. The valley or airshed is considered a giant room ventilated through various windows and doors (see Figure 2.33). The walls of the room are the Cascade Range in the east and the Coastal range in the west. The room is approximately 200-km long (N-S) and 50-km wide (E-W) and is partially closed at the north by rising hills and at the south by the Klamath mountain range. To the east, the average height of the wall is 1700 m, with only one major door, the Columbia River Gorge at the upper right. Minor windows exist along the east wall in the form of mountain passes. The west wall reaches an average height of 1000 m with one major break along the Columbia River and three minor passes, as illustrated in the figure by the door at the upper left and the three windows.

The artificial ceiling of the room is determined by the mixing height, which is the upper limit of active vertical mixing of the valley air. This height is primarily a functon of the vertical temperature distribution or the

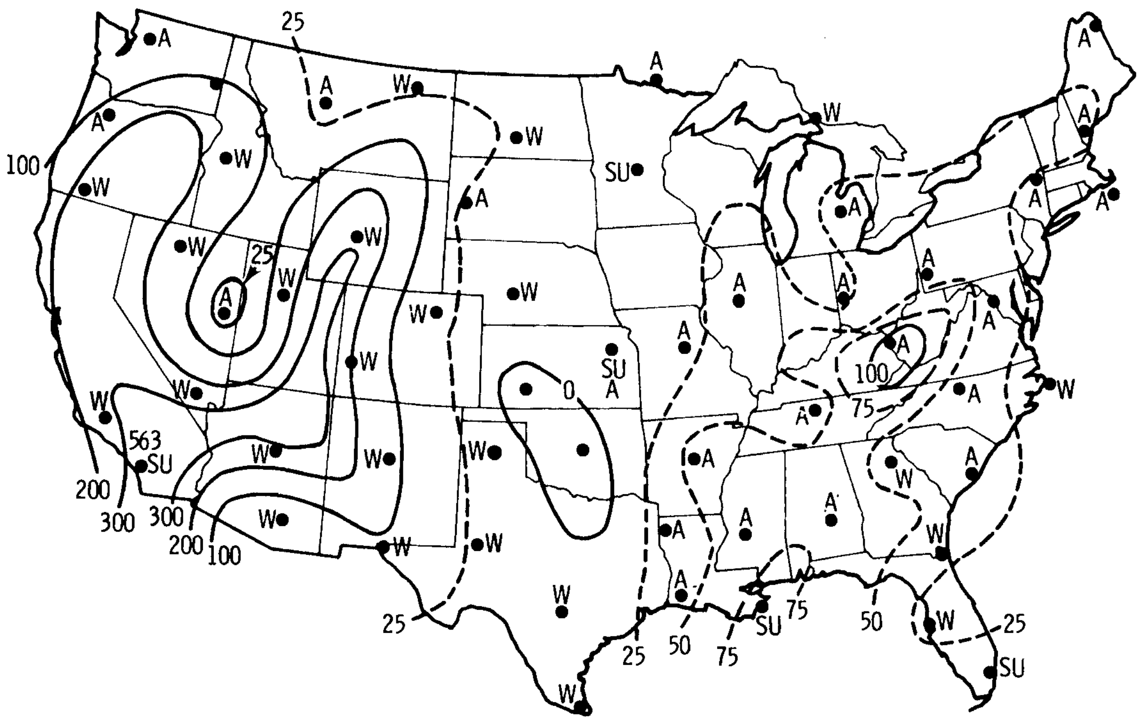


FIGURE 2.32. Isopleths of Total Number of Episode-Days of Limited Dispersion in 5 Years. Season in Which Most Episode-days Occurred at Each Station Indicated as W (Winter), SU (Summer), SP (Spring), or A (Autumn).

lapse rate and varies from nearly zero on clear nights to several thousand meters on days with good mixing. periods of stagnation are usually the result of persistent high pressure over the region during late summer and early fall when the mixing height can remain less than 1000 m for a week or more. During such periods, sea breezes and valley-slope winds are of major importance in ventilating the room through the doors and windows.

Bell and Thompson (1980) examined the effect of cross valley winds on ventilation of valleys. Numerical and laboratory experiments indicated that a modified Froude number, $F = \bar{U}/Nh$, is a critical parameter. In this case, $\bar{U}$ is the mean horizontal velocity above the crest of the valley sides: N is the Brunt-Vaisala frequency; h is the height of the valley walls. Ventilation occurs whenever F exceeds a value of 1.3; stagnation when F is less than 1.3 provided that the ratio of the total depth to the crest to trough depth (D/h) is greater than about 1.8.

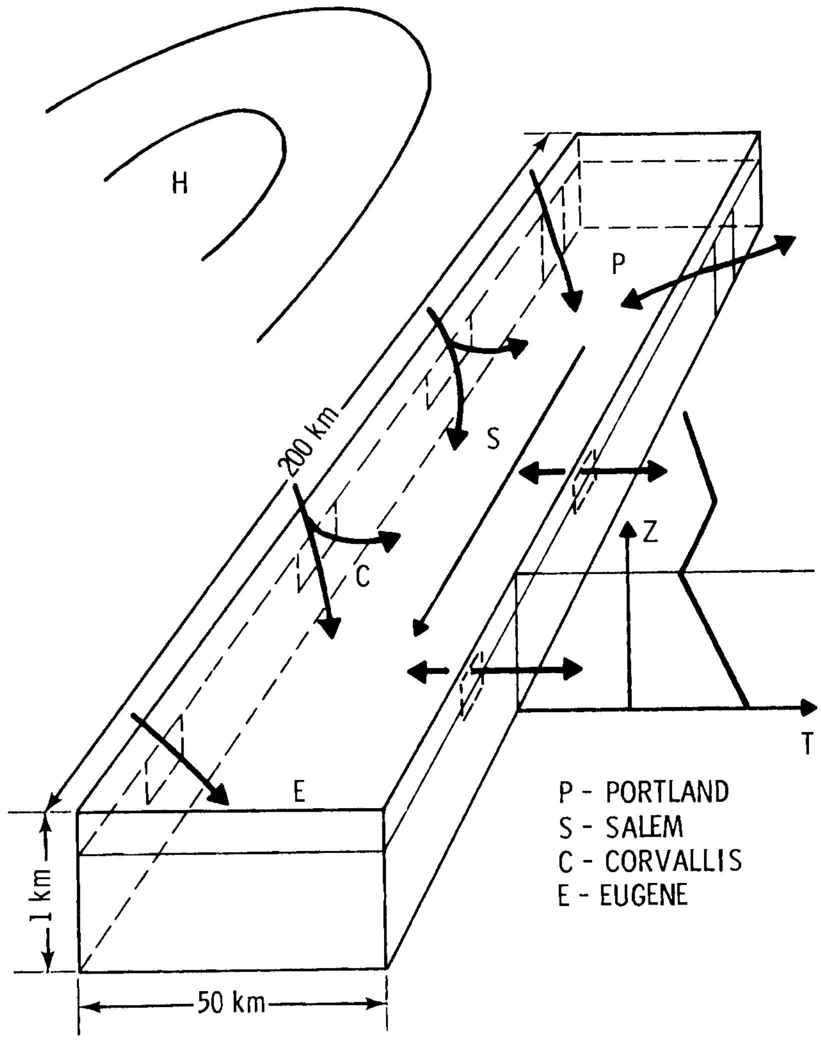


FIGURE 2.33. Schematic Diagram for Illustrating Natural Ventilation of the Columbia-Willamette Valley Airshed. The flow pattern illustrates typical summer conditions (adapted from Olsson and Peterson 1971).

2.5.2 Basins (Open and Closed)

Some examples of open basins are the Los Angeles, Columbia River, Snake River, and the Colorado-Green River basins. All experience stagnation episodes. Blocking mountain ranges to the east and a shallow mixing layer caused by the persistent elevated subtropical temperature inversion promote the poor natural ventilation of the Los Angeles Basin. Except for occasional fronts or storms the majority of the natural ventilation of the region is the result of monsoon sea breeze on-shore circulation and valley-slope wind systems (DeMarrais, Anderson and Cramer 1965; Edinger 1959 and 1963; Edinger et al. 1972; and Blumenthal, White and Smith 1978). Stagnation and poor natural ventilation episodes will certainly play a role in the siting and environmental evaluation of coal-fired power plants and the development of other energy sources in the Colorado-Green River Basin (Bowers, Anderson, and Cramer 1978; Johnson et al. 1977; Hinman and Leonard 1977a,b; and Topham 1979). Such episodes have been a pollution problem in the East for several years (Reisinger and Crawford 1979).

A closed basin or valley suggests that its surrounding ridges encircle the enclosed region completely except for windows in the form of passes or gaps. Naturally, this type of basin is ideal for stagnation since its surrounding ridges reduce the ventilation effects from synoptic storms; thus, a good share of the natural ventilation would have to be due to the local thermal-slope wind systems.

2.6 EDDIES AND VORTICES

As an airstream moves over or into inhomogenous terrain, it is subject to thermal and dynamic forces that tend to cause the airstream to develop vorticity or eddies. Generally, this vorticity occurs in four modes: 1) transverse (approximately perpendicular to the wind direction), 2) longitudinal (approximately in the same direction as the wind), 3) quasi-horizontal (Karman vortex street), and 4) vertical concentrated vortices.

2.6.1 Transverse or Separation Eddies

Transverse or separation eddies are the direct result of separation and wakes as discussed in Sections 2.2 and 2.3. Much remains to be known about the characteristics, lifetime, and dimensions of separation eddies as developed from different landforms.

2.6.2 Longitudinal Vortices

Helical rotors or vortices are the product of different combinations of the upper-level gradient wind speed and direction, the orientation of the valley axis with respect to the wind direction, the size of the valley, height of ridgeline and heating and cooling of the valley slopes. Wanta and Lowry (1976) present the general features of some possible combinations of mechanical and thermal perturbations in a simple ridge-valley system (Table 2.2 and Figure 2.34). Their four basic forms of airflow depicted in Figure 2.34 may be

TABLE 2.2. Generalized Mesoscale Windflow Patterns Associated with Different Combinations of Wind Direction and Ridgeline Orientation

Wind Direction Relative to Ridge	Ridge Orientation	Time	Resulting Flow
Parallel	East-West	Day	South-facing slope is heated. Single Helix.
Parallel	East-West	Night	Downslope flow on both slopes. Double Helix.
Parallel	North-South	Day	Upslope flow on both heated slopes. Double Helix.
Parallel	North-South	Night	Downslope flow on both slopes. Double Helix.
Perpendicular	East-West	Day	South-facing slope is heated. North wind--stationary eddy fills valley. South wind--eddy suppressed, flow without separation.
Perpendicular	East-West	Night	Indefinite flow--extreme stagnation in valley bottom.
Perpendicular	North-South	Day	Upslope flow on both heated slopes--stationary eddy one-half of the valley.
Perpendicular	North-South	Night	Indefinite flow--extreme stagnation in valley bottom.

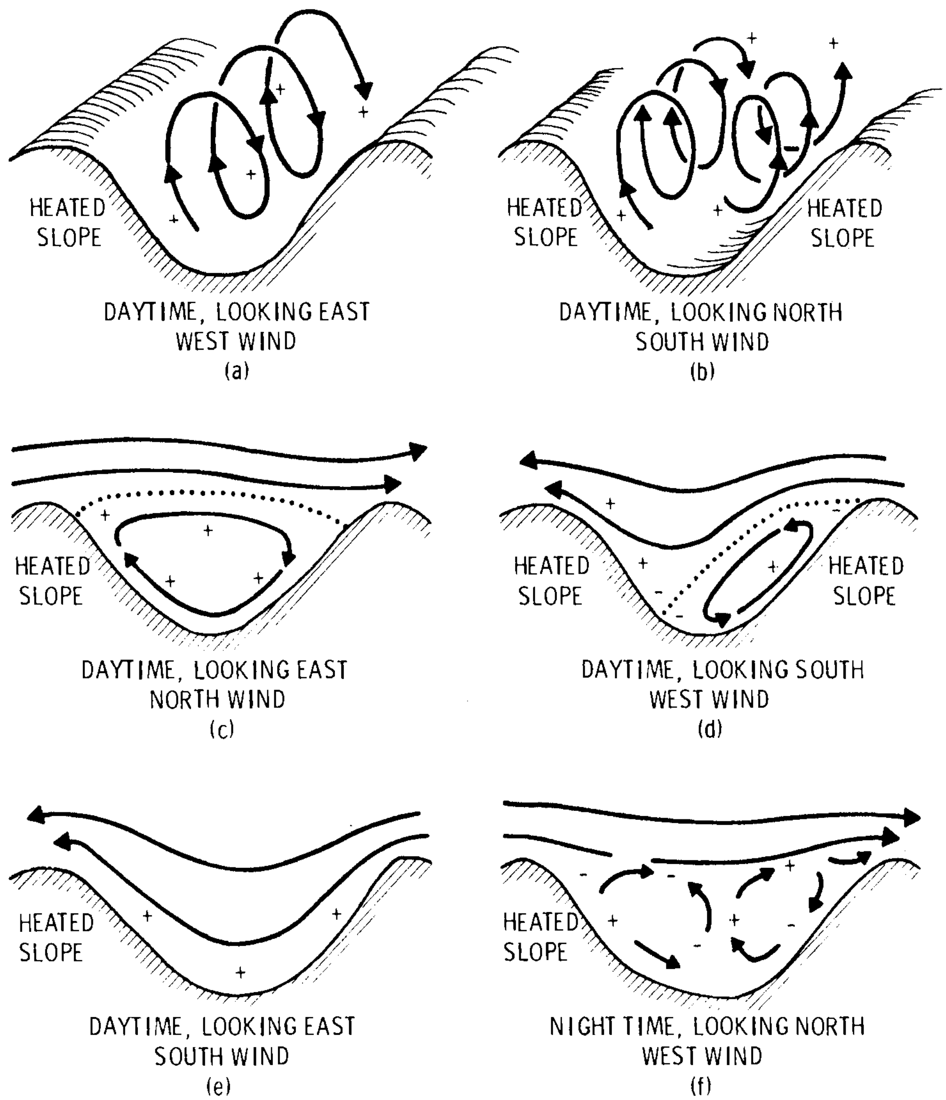


FIGURE 2.34. Mesoscale Flow Patterns that Could Develop in a Heated Valley as Described in Table 2.2. Reinforcement of thermal and mechanical circulations are marked +; opposition, - (adapted from Wanta and Lowry 1976)

roughly classified as: a) and b) helical, c) and d) isolated eddy, e) quasi-laminer, and f) inaefinite. Helical flows may be single or double, and isolated eddies may fill all or only part of the valley.

If a valley runs north to south, the east-facing slope will show upslope winds for some hours after sunrise. With a valley wind already established, the trajectories will be spirals at least in part; they will go up the east slope and proximally over to the west slope and down to the bottom again (Buettner and Thyer 1966, and Hewson and Gill 1944).

Strong wind speed and directional shear in the upper portions of the valley may also generate helical rotors. Observations taken with dual constant-volume balloons in the Eagle River Valley in central Colorado showed the presence of an elevated cross-valley helical rotor system during a period of strong (>20 ms^{-1}) upper-level cross-valley winds; surface winds in the valley were generally below 4 ms^{-1} (Wooldridge and Orgill 1978; Orgill 1971).

2.6.3 Quasi-Horizontal Vortices

The movement of stratified airflow over and around three-dimensional obstacles such as isolated hills, mountains, and islands can generate vortex shedding. This phenomenon has been demonstrated in laboratory flow streams as well as in the physical world. An experimental study was conducted by Brighton (1978) on the shear flow of stratified fluid around and over a simple conical obstacle when the Froude number ($Fr = U/Nh$) was much smaller than one. The fluid streams were found to be in nearly horizontal planes except near the top

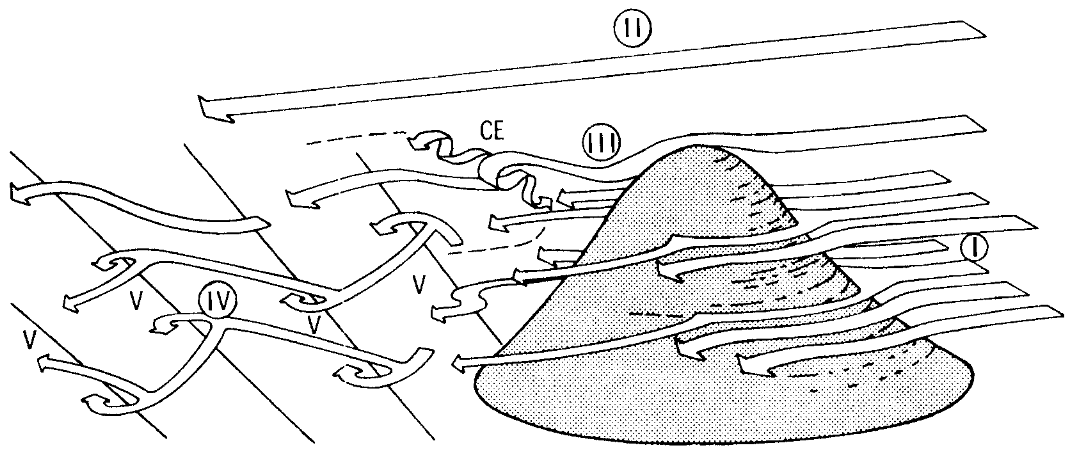


FIGURE 2.35. Three-Dimensional Illustration of the Most Prominent Features of Strongly Stratified Flow Past an Idealized Hill; Region 1: Potential Flow, Region 2: Flow Unaffected by Obstacle, Region 3: Development of Lee Waves and Sometimes a Cowhorn Eddy (CE), Regions 4 and 5: Wake Flow and Vortex Shedding (adapted from Brighton 1978)

of the obstacle. In the lee of the obstacle at the top of the obstacle, a cowhorn-shaped eddy with horizontal axis was observed; below this level, there is turbulent wake flow. Vortices are shed provided the Reynolds number is large enough and provided Fr is less than about 0.15. The shedding frequency was the same at all heights (Figure 2.35).

The first TIROS weather satellites of the early sixties and later satellites have shown the existence of eddy patterns in the wakes of large islands, such as the Azores, Canaries, Hawaii, and Madeira (Chopra and Hubert 1964 and 1965a,b; and Zimmerman 1969). These quasi-horizontal eddies are made visible by patterns in stratocumulus clouds lying beneath a strong temperature inversion about 0.5 to 2.0 km above the ocean surface and have been observed to persist for about a thousand kilometers downstream. The band width of the eddies is of the order of the cross-stream diameter of the island. Under favorable conditions, the pattern of eddies bears a strong resemblance to the classical Karman vortex street pattern observed in laboratory studies of wakes behind obstacles (Barnett 1972).

An analysis by Chopra and Hubert (1964, 1965a,b) indicates that viscosity (friction) plays two roles in the downstream eddy pattern. First it leads to the formation of the boundary layer in which vorticity is generated and the vortex pairs are formed, the vortices being shed alternately near each edge of the island (Figure 2.36). Second, the strength of the eddies is dissipated by diffusion; i.e., the size of an eddy increases and its strength decreases as it propagates downstream until its region of influence overlaps that of a neighboring eddy of opposite circulation.

The rate N of shedding of the eddy pair is given by

$$N = \frac{U_e}{a} = \left(\frac{1}{a}\right)\left\{U_o - \frac{K}{2a} \tan h \left(\frac{\pi h}{a}\right)\right\} \tag{2.19}$$

where U_e is the speed of propagation downstream of the eddies in a coordinate system fixed to the obstacle, U_o is the speed of the undisturbed flow; a is the longitudinal spacing between two eddies of similar circulation of strength K, and h is the lateral spacing (band width) between the two vortex rows. Chopra

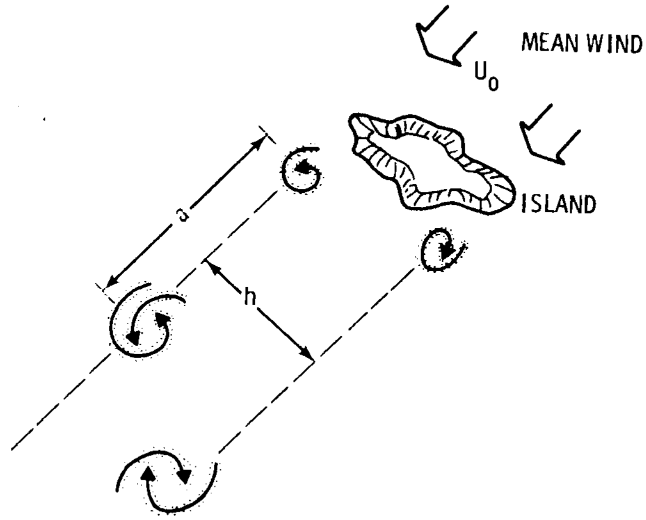


FIGURE 2.36. Schematic Illustration of a Karman Vortex Street in the Wake of an Island. Symbols a, h, and U_0 explained in text

and Hubert (1965) used satellite data to estimate the rate of shedding of eddies as

$$N = \frac{U_e}{a} = 0.13\ \mathrm{hr}^{-1} \qquad U_e \sim 7.1\ \mathrm{ms}^{-1} \quad a \sim 190\ \mathrm{km} \tag{2.20}$$

The lifetime of an eddy was estimated as

$$t = L/U_e \sim 21\ \mathrm{hr} \tag{2.21}$$

where L is the length of the vortex street. Zimmerman (1969) estimated the lifetime of an eddy as 34 hr, its maximum tangential velocity at a radius of 35 km at 3 ms^{-1} and its rate of kinetic energy dissipation as 0.2 erg gm^{-1} sec^{-1}. The effective eddy viscosity of a vortex pair as $2 \times 10^7\ \mathrm{cm}^2\ \mathrm{sec}^{-1}$.

2.6.4 Vertical Concentrated Vortices

Actual observations of vertical concentrated vortices in complex landforms are normally quite rare and their contribution to transport and dispersion is probably relatively minor. However, these wind systems (dust devils, tornadoes, fire whirlwinds, mountainadoes) are briefly included since topography may play a role in the development or nondevelopment of these type of vortices.

The common dust devil is a small (diameter ~1 to 30 m) but vigorous vortex created during hot days over dry terrain.[a] Superimposed on its strong rotary motion is a significant vertical motion that results in a combined flow pattern similar to that of a helical vortex. Dust devils do not depend upon cloud systems or moisture as a source of energy for their genesis and are normally a fair weather phenomenon; essentially, they are thermal convection elements.

A dust-devil-like vortex has been observed on lee slopes with separation (Scorer 1978 and 1972). If airstreams from two different sides of a hill or ridge have different components of velocity along a line of separation, the upward acceleration from the line may stretch vortex lines that result in individual vortices. Dust whirls and whirls of blown snow often occur on the lee slopes and sometimes pieces of vegetation are carried into the air by the vortex. These vortices are usually short lived.

Mountain thunderstorms are fairly common in complex landforms during the summer season. However, the observations of funnel clouds or tornadoes over regions of complex terrain are rather rare (Feris 1970 and Budney 1965, and Koscielski 1967). Whether this is caused by the absence of certain important antecedent conditions for vortex development over complex terrain or by the influence of topography has not been positively determined.

Fire whirlwinds, fire devils and fire tornadoes are strongly buoyant vortices of fire-heated air (Morton 1969). They often occur with large-scale forest fires or controlled fires. Observations suggest that topography such as the leeside of ridges may play a role in the development of fire vortices (Graham 1955).

Mountainado is an unofficial term defining concentrated vertical vortices that accompany strong wind storms (>20 ms^{-1}) on the lee side of the Rockies

(a) Actually, vortices of dust-devil size may occur over water, snow, land, or ice as long as the conditions for formation are satisfied.

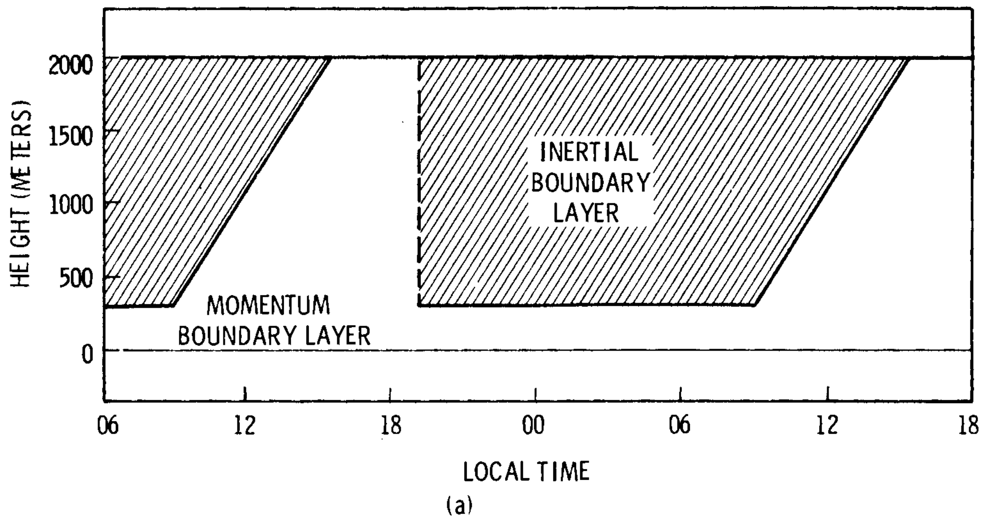
(Bergen 1976). The so-called mountainadoes have been observed in several different forms, including tall, narrow pillars and broad, rotating drums. The vertical ascent of entrained debris and definable vortex tops have also been observed. Rotational velocities are estimated at around 40 ms^{-1} and can cause considerable damage.

The development of mountainadoes has been postulated as being caused in part by vortex shedding from smaller topographic obstacles and to the vertical orientation of small horizontal eddies. Thermal forcing may provide some additional energy to prolong and expand the vortex, but is not seen as a generation mechanism because of the strong, turbulent nature of the downslope wind storms.

2.7 MOMENTUM AND THERMAL BOUNDARY LAYERS

The atmospheric boundary layer, i.e., the region of air below about 2000 m, has generally been subdivided into three layers: molecular, surface, and planetary or Ekman (Panofsky 1974). In the lowest millimeter or so of the boundary layer, molecular properties strongly influence the structure of temperature and other scalars, but not of momentum unless the ground is unusually smooth. Above the molecular boundary layer is the surface layer. This is the region, up to 30 m, where the vertical variation of stress and other fluxes can be neglected, and the wind direction is essentially constant. In this layer, stratification is important but the earth's rotation is not. Height of the layer is quite variable particularly at night when it may be only a few meters thick.

Above the surface layer is the planetary or Ekman layer in which stratification and rotation play important roles. Further division of the Ekman layer has been made by Panofsky (1974) and Hoxit (1975). Panofsky defines a tower layer that is generally below 150 m but the subdivision is not useful in nocturnally stable air because of significant turning of the wind. Hoxit subdivides the nocturnal Ekman layer into two sublayers. The layer adjacent to the surface is called the momentum boundary layer and is only a few hundred meters thick at night. In the daytime it corresponds to the Ekman layer. The second layer is formed shortly after sunset and eventually extends from around 200 to 1500 m. This layer is the inertial boundary layer (Figure 2.37a).



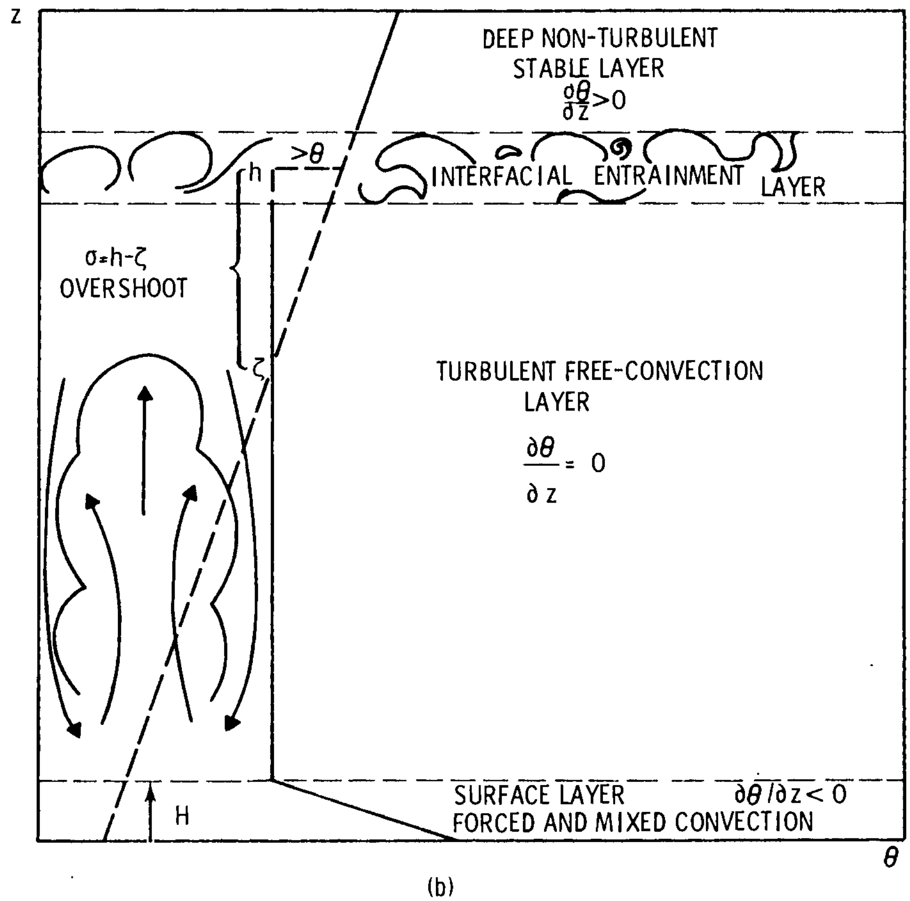


FIGURE 2.37. Types of Boundary Layers Over Land: a) Diurnal Variations in the Depths of the Momentum and Inertial Boundary Layers, and b) Developing Convectively Unstable Boundary Layer as Shown by Potential Temperature Profiles (adapted from Hoxit 1975 and Carson 1973)

During the daytime or storm periods another type of boundary layer may develop within the Ekman layer called the convective unstable boundary layer. Models of a convective unstable boundary layer capped by a stable layer have been proposed by Ball (1960), LaVoie, Cotton, and Hovermale (1970); Deardorff and Willis (1967); Deardorff (1972 and 1974); Lilly (1968); and Carson (1973).[a] Figure 2.37b gives a schematic representation of the potential temperature (θ) profile in a convectively unstable boundary layer. The model consists of four layers: 1) a shallow superadiabatic layer ($\leq$few tens of meters) with its large vertical shears of wind and temperature incorporates the regions of forced and mixed convection; 2) a turbulent free-convection layer is buoyancy dominated and the θ profile is virtually independent of height although it is often observed to be slightly stable, particularly in the upper region of the layer; and 3) an interfacial entrainment layer is an unstable-stable interface that has a highly contorted, almost undefinable surface, because of the physical overshooting into the stable layer of energetic convective elements originating in the surface layers and continually bombarding the interface. These layers vary in depth and character, and the temperature field indicates marked spatial and quasi-discontinuous variability; and 4) a deep non-turbulent stable layer caps the turbulent free-convection layer. Field measurements of potential temperature in rough terrain in Colorado during winter storms have exhibited convectively unstable boundary layers. Wind tunnel models have reproduced temperature profiles similar to a convective unstable boundary layer (Orgill, Cermak and Grant 1971, and Orgill and Cermak 1972).

The structure of boundary layer flows (profiles of mean wind, temperature, moisture, and turbulence) depends on the nature of the underlying surface as well as temperature (stability) and moisture. When airflow or other variables are required to readjust to a new set of boundary conditions, transitional zones or internal boundary layers develop downwind from the discontinuity. Generally, internal boundary layers fall into four categories (Munn 1966).

(a) See Plate (1971) for a review of the work on convective unstable boundary layers.

- Those that result from a discrete change of roughness z_0 (Momentum Internal Boundary Layers).
- Those that are produced by a barrier such as, a fence, hedge, line of tress, or small ridge.
- Those that are caused by change in surface temperature and moisture (Thermal Internal Bounaary Layers).
- Those that result from extending the surface temperature and moisture changes to cases in which the properties of the underlying surface are continually changing.

2.7.1 Momentum Internal Boundary Layers (MIBL)

Typically the boundary-layer flow alters over a sudden change in roughness from a surface whose roughness is z_0 to a surface with a roughness height of z_{01} to a rougher surface with a roughness height of z_{02}, as depicted in Figure 2.38a. Elliott (1958), Panofsky and Townsend (1964), Bradley (1968), and Rao, Ungaard, and Cote' (1974) have evaluated the affect of a change in roughness. (See Plate 1971 and Munn 1966 for surveys of past work.) Elliott (1958) has shown that the boundary layer grows according to

$$\delta \propto x^{0.8} \text{ for } x/Z_{02} > 10^3 \tag{2.22}$$

where δ is the internal boundary layer thickness and x is the downwind distance. This growth of the boundary layer is independent of wind speed but slightly dependent on thermal stability, raising the height of the layer in lapse conditions and lowering it in inversions.

The development of momentum internal boundary layers over a small two-dimensional obstruction is complex (see Figure 2.38b). According to Plate, no less than seven flow zones of different aerodynamic behavior can be distinguished. In relation to actual terrain effects Plate remarks:

> Any added complexity of the terrain increases the difficulty of analytical treatment while at the same time making the situation more special and of less general applicability. Very soon a point is

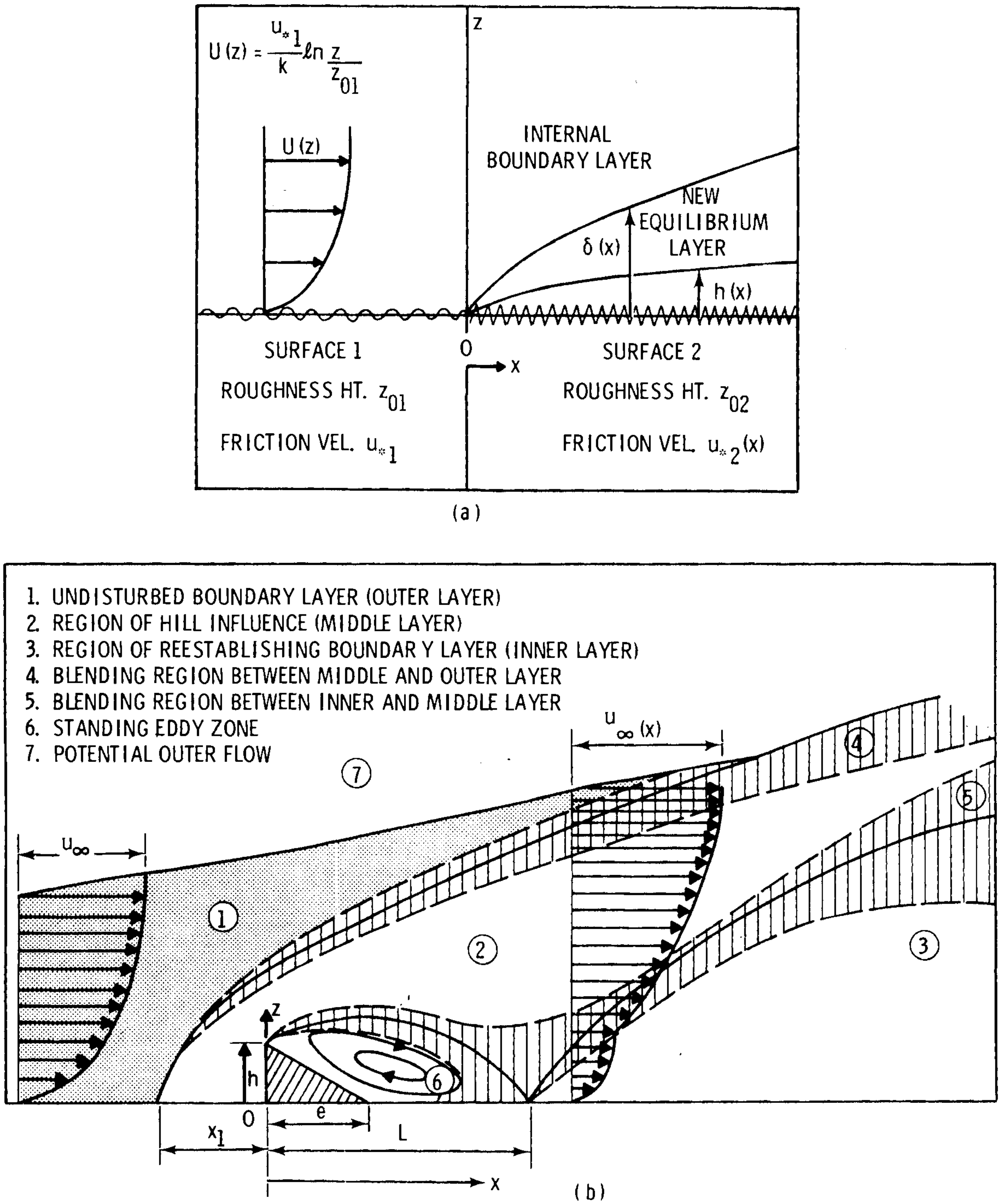


FIGURE 2.38. Development of Momentum Internal Boundary Layers: a) the Internal Boundary Layer (IBL) Over a Step-Change of Surface Roughness, and b) Flow Zones of a Boundary Layer Disturbed by a Shelterbelt or Topographic Obstacle (adapted from Plate 1971)

> reached when further investigations of more complex terrains become impractical; the amount of time and effort expended in solving the problem for the flow field is nolonger in reasonable proportion to the value of the information obtained. It is for these situations that field or wind-tunnel measurements on the actual terrain or a model thereof must be made (p. 177).

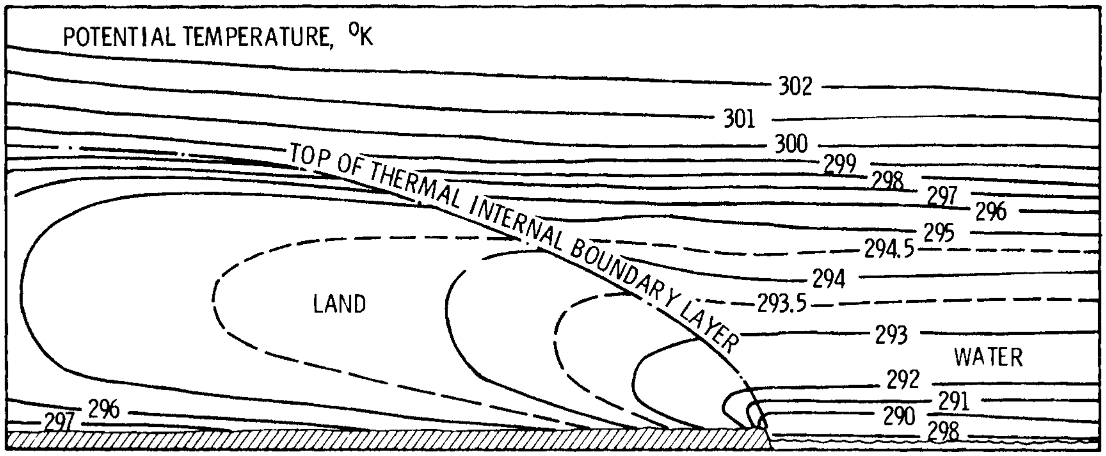
Hoecker and Angell (1969) illustrated the complexity of airflow moving from water to hilly coastal terrain. Their study showed that wind speeds at a height of 300 m were stronger over land than over sea, wind direction backed (counterclockwise) by 4° to 11° at heights 200 to 400 m, and the maximum upward velocity occurred 250 m above the sudden change in terrain height (~100 m) with compensating downward motion commencing 3 km inland.

2.7.2 Thermal Internal Boundary Layers (TIBL)

Thermal internal boundary layers develop when an oceanic marine layer flows onto a warm continental coastal area or island (Edinger 1959 and 1963; and Raynor, Setharaman, and Brown 1979) or a lake breeze moves onto a warm shoreline (Lyons 1975 and Moroz 1967). The interface is created since the two surfaces rarely have the same temperature and almost always differ in aerodynamic roughness. The interface typically starts at the surface discontinuity and slopes upward in the direction of the flow at a rate dependent on wind speed, the original characteristics of the air, and the properties of the downwind surface (see Figure 2.39a).

Coastal boundary layers may be caused by differences in surface temperatures, by differences in surface roughness, or by differences in both properties. Most situations occur when the land is rougher and warmer than water or it is rougher and colder than water. Cases with the two surfaces at the same temperature are rare and of brief duration, while cases with land smoother than water seldom exist.

A thermal internal boundary layer may form when stable air travels over warm terrain, such as a town or city in a mountain valley, and a mixed layer develops to a specific height (Figure 2.39b). The height of this mixing layer is an important variable in air pollution situations over towns or cities at night in enclosed valleys because of the restricted ventilation. Summers (1964) has developed a simple model for calculating the mixing depth height and



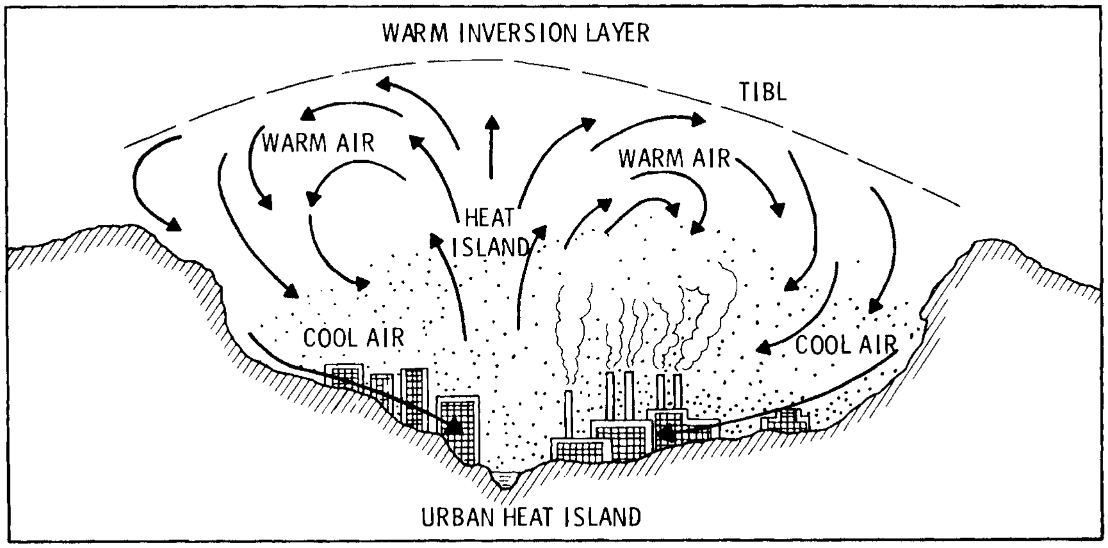


FIGURE 2.39. Types of Thermal Internal Boundary Layers: a) a Thermal Boundary Layer (TIBL) Developed Between Land and Water. Heavy line denotes the top of the layer heated by penetrative convection from the surfaces and b) a TIBL over an urban heat island

the heat-island temperature excess. Modifications would be required for application to valley situations because of terrain and boundary conditions.

2.7.3 Phenomena Associated with Momentum and Thermal Boundary Layers

Two phenomena which occur within momentum and thermal boundary layers that are of interest in complex terrain situations are low-level jets and nocturnal destablization.

2.7.3.1 Low-Level Jets

Examples of low-level wind maxima (LLWM) or jets may be found almost anywhere in the United States during any season of the year (Bonner 1968, and Blackadar 1957). Krishna (1968) studied the diurnal variations of winds as a function of latitude and found that low-level jets:

- occur before midnight in midlatitudes, slightly after midnight at 30°N, at sunrise at 17.5°N., and later farther south.
- increase the amplitude of the diurnal wind speed wave from north to south, reach a maximum a little below 30°N, and then decrease rapidly. The supergeostrophic winds are strongest between 40°N and 20°N suggesting that these latitudes are more favorable.
- have a height below 500 m north of 30°N, about 550 m between 30°N, and 12.5°N and higher farther south.

The phenomenon is best developed over the Great Plains in the United States during spring and summer nights. This wind variation is especially pronounced with southerly flow along the sloping terrain to the east of the Rocky Mountains. Boundary layer winds oscillate diurnally reaching a maximum speed at night and a minimum during the day at elevations around 1000 m above the ground. The amplitude of this oscillation is 2 to 5 ms^{-1}. Maximum speed occurs between 00 and 03 local time (Figure 2.40).

Explanations have been given for the development of the low-level nocturnal jet (Blackadar 1957, Wexler 1961, Holton 1967, and Lettau 1967). Blackadar and Buajitti (1957) indicate that the boundary-layer wind undergoes diurnal speed and directional variations as a result of the diurnal variation of frictional coupling of the wind with the ground. The forces involved are shown in Equation (2.23) and Figure 2.41.

$$\dot{V}/ = -\alpha\Omega p - 2\Omega \times V/ + \alpha/F \qquad (2.23)$$

During the afternoon the pressure gradient, Coriolis, and frictional forces are in quasi-balance and the boundary-layer wind does not accelerate. When friction decreases in the late afternoon and evening, the forces become

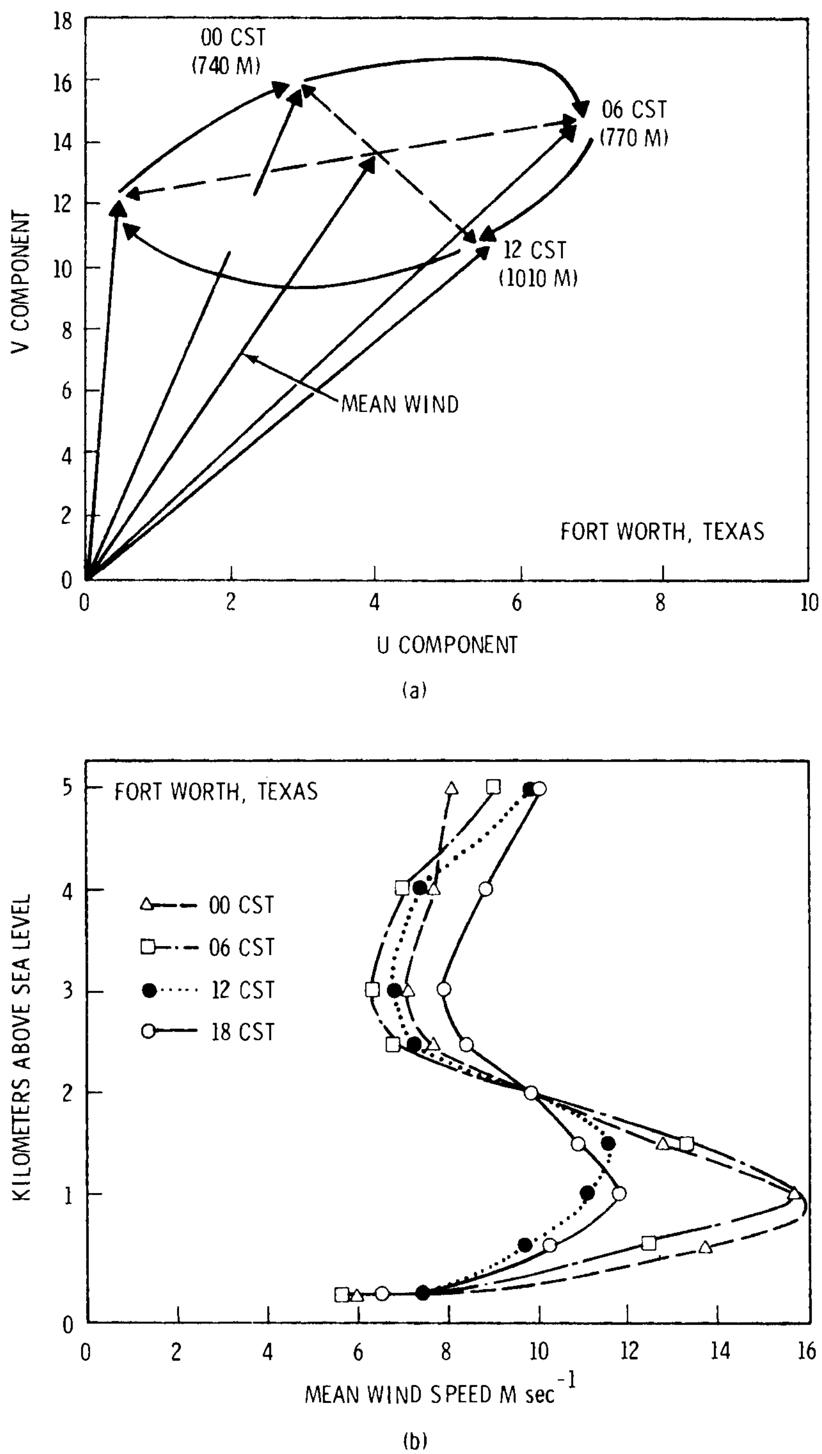


FIGURE 2.40. Development of the Low-Level Nocturnal "Jet": a) Vector Mean Winds on 16 Summer Days. Mean Altitudes in Meters Above the Ground at Each Observation Time are Indicated in Parentheses (note the rotation about the mean wind) and b) Mean Wind Speeds Profiles for Same 16 Summer Days (adaptedfrom Bonner 1968)

unbalanced and the real wind accelerates in a direction shown by vector $\dot{V}$/ in Figure 2.41. The wind may then undergo an inertial oscillation with a period of about one day and the real wind may become supergeostrophic some time during the night. This mechanism duplicates the observed oscillation but does not consider the effect of sloping terrain. Holton (1967), Lettau (1967), and Bonner and Paegle (1970) have developed models that incorporate the effects of sloping terrain. These models have approximated the observed oscillation in the lower winds.

Information on the existence and variation of low-level jets in complex landforms is sketchy. Barad (1961) suggests that they are least likely to be found in hilly or mountainous areas. Yet, Hoecker (1965) in an investigation of the nocturnal jets over the Great Plains found that the jetlike winds were evident over the hilly terrain of Arkansas but were weaker and higher (500 to 700 m above ground) than the Great Plains counterpart. Hoecker suggests that the increased frictional drag with the terrain caused the jets to increase in height.

Bonner (1968) in his study ignored stations within the Rocky Mountains because of their high elevations and the expected local effects caused by

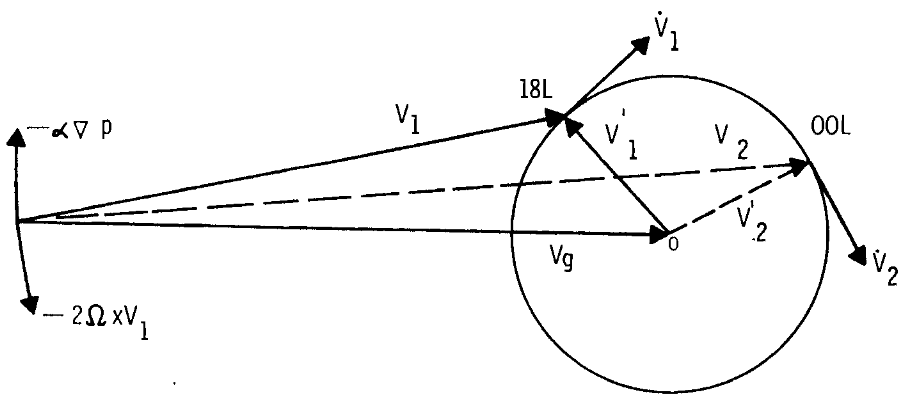


FIGURE 2.41. Schematic Representation of the Effect of the Inertial Oscillation on the Real Wind in the Boundary Layer. V is the rotating acceleration vector and V' is the geostrophic departure vector (adapted from Bonner 1968)

mountain-valley wind systems. However, the existence of jets was noted for four West coast stations. Frenzel (1962) also noted the existence of jets in the central valleys of California.

Wind data collected at White Sands Missile Range (Tularosa Basin), New Mexico, frequently show the existence of nocturnal jets and have been classified into three different types (Rider 1966, and Rider and Armendariz 1971). Type I is a northerly maximum wind generally found at heights of 100 to 400 m and occurring on nights when there is a high pressure system over the central and northern Rockies with mostly clear skies and decreasing geostrophic winds with height. Type II is a west-northwest or northwest jet occurring as a cold trough aloft moves eastward through the area during the night after a surface temperature inversion has developed. Timing of these conditions appears to be important for development. The height is between 600 and 1000 m. Type III is found with southerly winds and reaches its greatest development from 0400 to 0600 local time at a height of 400 to 600 m. This type appears to be similar to the nocturnal jets observed over the plains just east of the Tularosa Basin.

Significant nocturnal jets occur on occasion in the Colorado - Green River Basin (Rider and Armendariz 1966). Limited data suggest that the phenomenon is less common and the shear and wind speeds are less than those observed in the Great Plains. The height of the maxima is a little higher and significant low-level maxima occur with lapse temperature conditions. The jets are most likely to develop on cloudless nights when there is a significant sustained pressure gradient force.

Additional evidence for the existence of jets in complex terrain is provided by Byram (1954, 1959) who points out the importance of wind profiles with jets in the blow-up phenomenon of forest fires, i.e., the sudden intensification of burning and rapid spread of a fire because of a dramatic change in the fire's convective column.

2.7.3.2 Nocturnal Destabilization

Wanta (1969) found that a characteristic antecedent feature of statically stable layers within the atmospheric boundary layer following sunrise is the abrupt onset of turbulence. The critical Richardson number for such transitions in the 46- to 125-m layer at Upton, New York, was determined to have a value of around 1/2. The typical sequence of events in a stable layer undergoing shear destabilizations is an increase in static stability and an increase in wind shear, followed by a breakdown of the stable configurations when both stability and wind shear decrease. The process is presumably iterated at increasing heights.

The destabilization process apparently operates at times other than following sunrise. During the early hours before sunset when stabilizing forces in the boundary layer are on the ascendant and perturbations in the flow are being damped a sudden burst of turbulence may occur. Pulsation of the surface wind on some clear windy nights may be another example.

Nocturnal destabilization may also occur with the development or dissipation of the nocturnal low-level jets. Blackadar (1957) indicates that the character of the wind profile is important in determining whether the growth of the nocturnal inversion is orderly and controlled or whether it is chaotic or perhaps entirely absent. When a wind maximum at the top of the inversion exists, the generation of turbulence within the inversion is under control and this prevents a chaotic breakdown of the inversion. However, if the wind maximum is above the level of the inversion, a chaotic breakdown of the inversion may occur. Blackadar's theory indicates that while there may be a tendency for the wind and temperature profile to adopt the more stable configuration, it is not necessary that they do. Hoecker (1963) and Bonner (1968) have shown that strong jet maxima can exist even with nearly adiabatic lapse rates.

2.7.4 Summary

Internal boundary layers have been observed and estimated over relatively homogeneous terrain for several years. Analytical, numerical and physical modeling of this phenomena has been relatively successful but models are usually limited to two dimensions.

Information on internal boundary layers over complex landforms is severely lacking. However, such phenomena may be very difficult to observe and measure in a mountainous environment where roughness exists on several different scales and the wind and temperature patterns can be very complex. Nevertheless, future field and modeling work should try to determine whether it is possible to define such phenomena as internal boundary layers and low-level jets in complex terrain.

2.8 TURBULENCE

Turbulence is the very disorganized motion of fluid or air in which it is not possible to predict the individual fluctuations from past history. Turbulence is therefore a statistically random process. In relation to the atmosphere, what constitutes turbulence has often been discussed. The present view is that a mean wind cannot be defined without specifying the time over which the average is obtained. The mean wind may be the average for a minute, an hour, a day, or a year; fluctuations about the mean are then called turbulence. Some common definitions are:

- The intensity or level of turbulence is given by the ratio $\left(\overline{U'^2_i}\right)^2$.
- Intermittency is the changing character of turbulence from one period to the next.
- Stationarity occurs when the probability distribution of the fluctuations is independent of time, i.e., there are no trends.
- Homogeneity occurs when the probability distribution is independent of position in the fluid. Turbulence may be homogeneous in any horizontal plane but not in the vertical direction because of shear.
- Isotropy is defined by the relation $\bar{U}'^2 = \bar{V}'^2 = \bar{W}'^2$.
- Spectrum is an analysis that determines the contributions of various frequencies to the total turbulent energy.
- Cascade pertains to the shape of the spectrum as given by Kolmogorov hypothesis: Energy enters the spectrum at relatively low frequencies and is transferred to higher and higher frequencies until it is finally dissipated.

2.8.1 Turbulent Intensities

The turbulent level or intensity of airflow over complex landforms such as ridges, hills, and valleys is obviously different than that over level ground. Smith and Wolf (1963) made a series of airflow and turbulence measurements over four widely differing landforms, two of which were in the hilly terrain of Nevada and Washington. Results indicated that the major terrain influence on the distribution of turbulent energy occurs with height; i.e., higher relief (hilly) regions maintain nearly constant turbulent energies to greater heights (~600 m) as compared to low relief regions. This energy appears to be maintained to a height roughly comparable with the relief height (see Figure 2.42) Grain--the dominant distance between principal terrain features--also played a role in the turbulence generation but appears to be of secondary importance relative to the relief.

The spatial distribution of atmospheric turbulence was measured near the eastern edge of the Deschutes River Canyon in Oregon (Horst and Powell 1974). For winds blowing from the west onto the upland from the canyon, the mean wind speed, the variance of the vertical component of the wind (σ_w^2), the total energy per unit mass (q^2), and the standard deviations of the elevation angle (σ_ϕ) were all found to be significantly higher near the canyon edge. The turbulent intensity of the wind at locations furthest from the canyon edge was found to be slightly higher under conditions of unstable and neutral thermal stability than found at horizontally homogeneous sites during similar conditions. Values more than twice as great as those at homogeneous sites were measured at locations furthest from the canyon edge during stable conditions.

Turbulent intensity profiles are expected to be more complicated and variable within ridge and valley terrain. During periods when the upper level wind crosses a valley axis obliquely, the region between intervalley and extravalley may exhibit high turbulent intensity. Constant-volume balloon data from central Colorado valleys indicate a maximum of turbulent vertical velocities at or near the level of the highest peaks (4400 m msl). A striking aspect of the vertical motion distribution was the gradual increase to that height (see Figure 2.43). Table 2.3 shows an estimate of the distribution in turbulent intensity with height from dual constant volume balloon flights on such a day

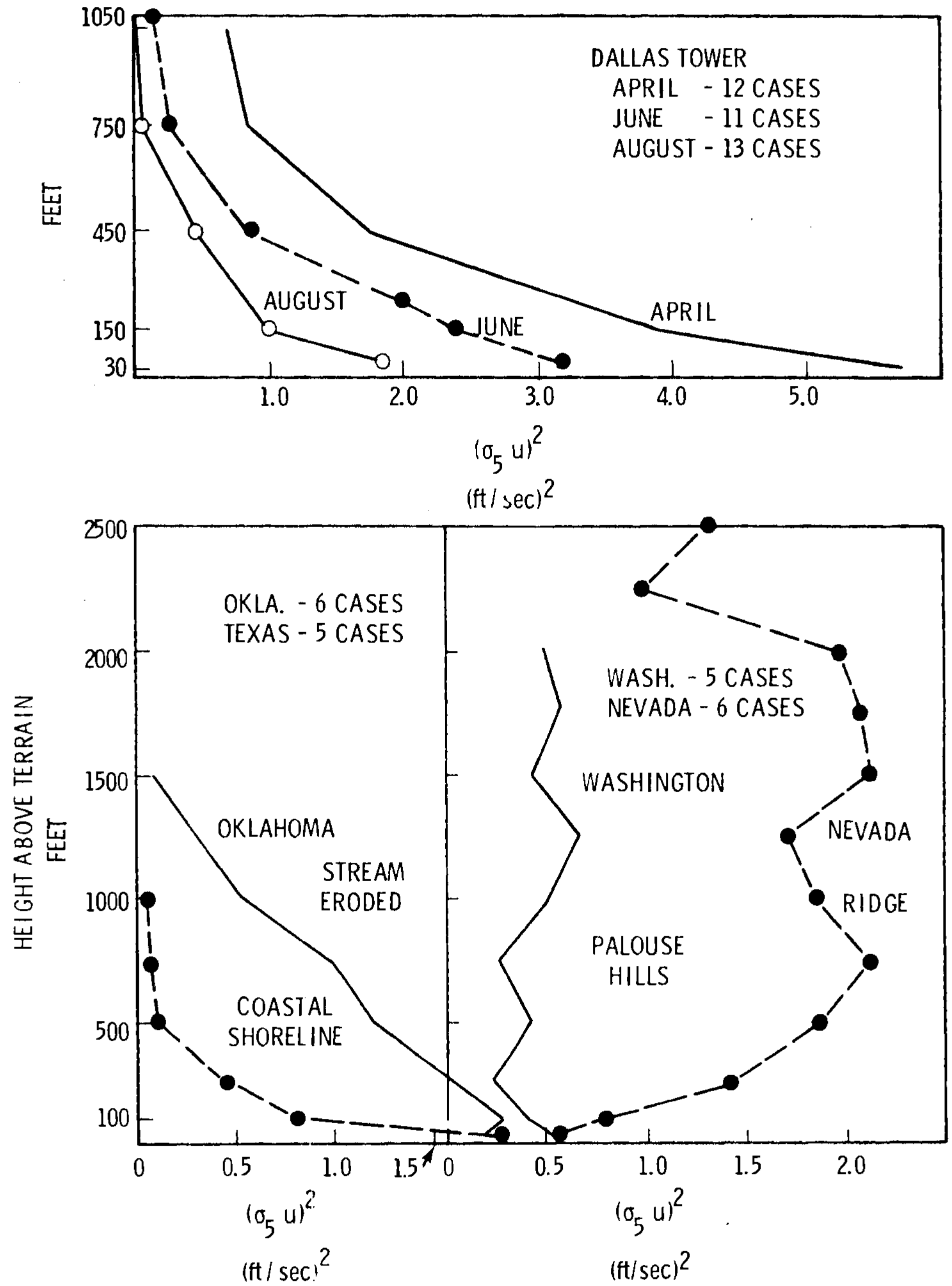


FIGURE 2.42. Turbulent Energy Profiles for Different Types of Landforms Varying from Relatively Shallow-Terrain to Hills and Ridges. $\varepsilon^{1/3} = 0.181\ \sigma_5\ U_0\ ^{2/3}$ where ε = turbulent dissipation rate, σ_5 = degrees and U_0 = ambient wind velocity (adapted from Smith and Wolf 1963)

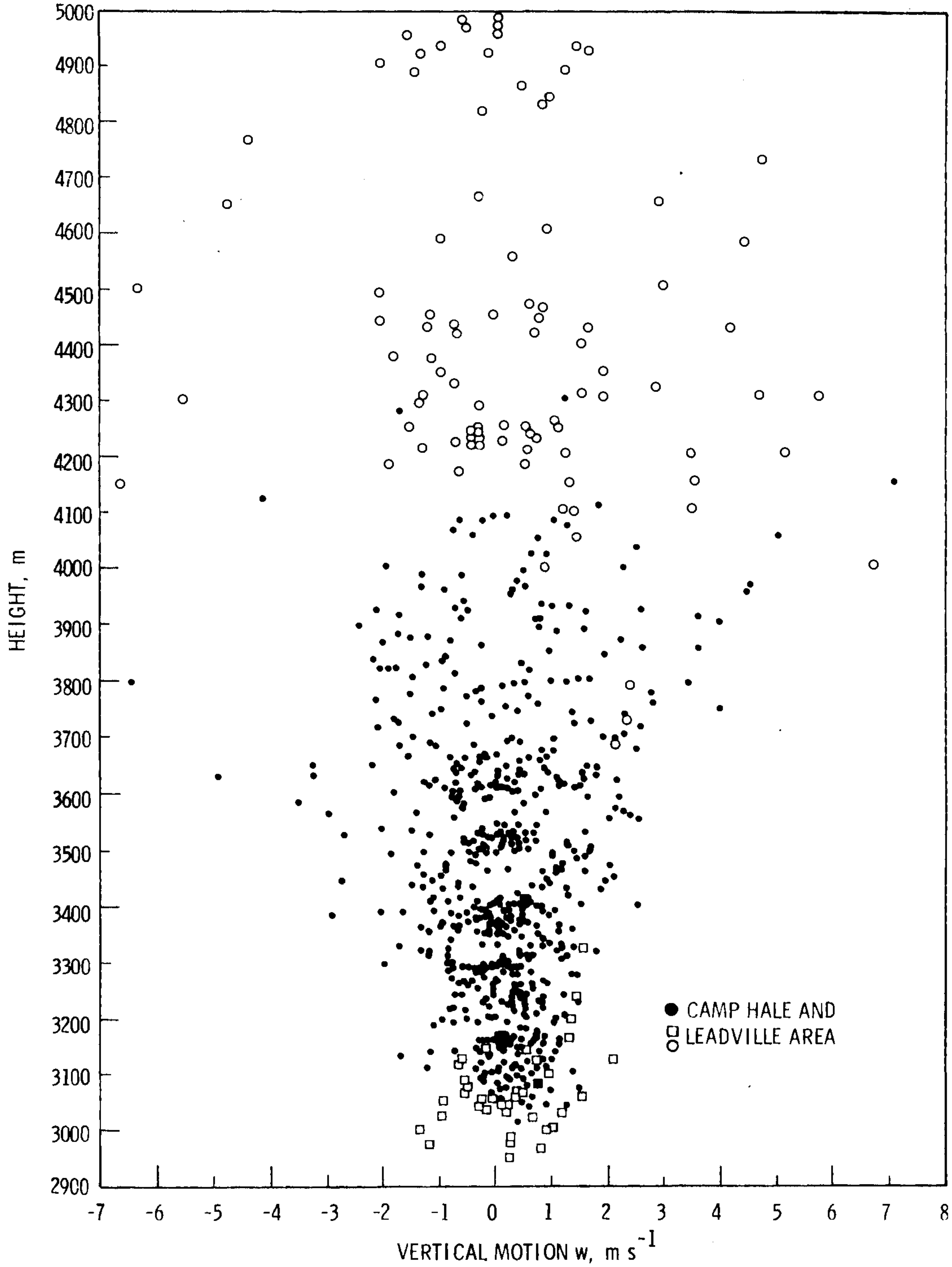


FIGURE 2.43. Vertical distribution of vertical motion with height over the Camp Hale (●) and Leadville (o,[]) areas. Data is derived from 16 superpressured balloon flights targeted for Chicago Ridge (from Wooldridge and Orgill 1978)

TABLE 2.3. Turbulent Intensity ($\sigma_i/\bar{U}$) Variations with Height for Dual Balloon Flights Within Eagle River Valley, Colorado, December 12, 1969

Velocity Component	Mean Elevation (km msl) 3.11	3.22	3.34	3.55	3.66
i = v (lateral)	0.10	0.21	0.28	0.84	1.32
i = u (longitudinal)	0.16	0.49	0.35	0.65	0.67
i = w (vertical)	0.12	0.13	0.19	0.50	0.74

(Wooldridge and Orgill 1978). Turbulence measurements were obtained by research aircraft for a valley, an isolated mountain (Elk Mountain) and other landforms in southern Wyoming by Connell (1976), Dawson and Marwitz (1979), and Karacostas and Marwitz (1980). Hydraulic jump phenomena, turbulent kinetic energy budgets, and other turbulent features of the airflow were investigated. Results from these studies show that shear production, eddy dissipation rate, buoyancy and vertical transport terms were important in the turbulent kinetic energy budgets. Shear production and eddy dissipation rate were dominant terms. The imbalance term was estimated to be relatively small.

2.8.2 Homogeneity, Stationarity, and Isotropy

Virtually all studies of atmospheric turbulence assume homogeneous and/or stationary turbulence. The concepts are rather restrictive, however, when investigating airflow in complex terrain. Horizontal inhomogeneities and the unsteadiness resulting from diurnal variations and other factors make it extremely unlikely that a quasi-steady approach will yield satisfactory results in describing turbulence. In many cases, the assumption of homogeneity and stationarity will have to be relaxed except on a very local basis (Lang and Hansen 1966). For example, stationarity of turbulent atmospheric flow was investigated by Wooldridge (1974) and Wooldridge and Ellis (1975) in the Cache Valley of northeastern Utah. Indicators for stationarity of mesoscale windpatterns were means, variance, turbulent intensities, and autocorrelograms of the components of the Lagrangian velocities. Data from four sequences of superpressure balloon flights were used in the analysis. Results indicated that the horizontal components of the Lagrangian velocities at levels below the ridges were only weakly stationary but the vertical component showed the highest degree of stationarity. Above the ridge levels, all components of the velocities exhibit reasonable stationarity in the turbulent flow.

Isotropic turbulence is not found in the lower atmosphere even under conditions of homogeneity and stationarity. It's unlikely that isotropic turbulence will be observed very frequently in airstreams over complex landforms.

2.8.3 Turbulence Spectrum

The turbulent dispersion characteristics of the atmosphere are visualized as a continuous cascade of eddies of ever decreasing size and along which turbulent energy is brought down from the largest synoptic energy input scale to the smallest scale where viscous dissipation occurs. Kinetic energy spectra display this cascade in terms of the turbulent kinetic energy of the wind as a function of either period, wave length or wave number. Other spectra may be wind speed and temperature. Because of the complex nature of the atmosphere, all these spectra display maximum and minimum as shown in Figure 2.44a.

An interesting feature that has emerged from spectra on horizontal velocity is the existence of a spectral gap in the micrometeorological range for velocity fluctuations with wavelengths on the order of 10 km and for time periods of 1 hour. In the mesometeorological range, Van der Hoven (1957) has confirmed the existence of a gap between turbulence of a-greater-than-5-hour-and-less-than-a-0.1-hour periodicity and has suggested that this is possibly due to the absence of a physical process that can support eddy energy in this range. Tyson (1968) suggests that the nocturnal downvalley mountain wind fills this gap. He has analyzed the horizontal velocity spectra of surges in the mountain wind in valleys of South Africa and indicates that the maximum turbulent energy is generated by waves of the order of 10 km in length for time periods of 1 hour, thereby filling the spectral gap (Figure 2.44b).[a]

(a) Tower wind data taken during July 1979 ASCOT experiments and at other time periods in the Anderson Creek Valley of the Geysers KGRA also shows that the spectral gap is filled by the local drainage or mountain wind. Daytime upvalley winds do not fill the gap (see Horst and Doran (1981)).

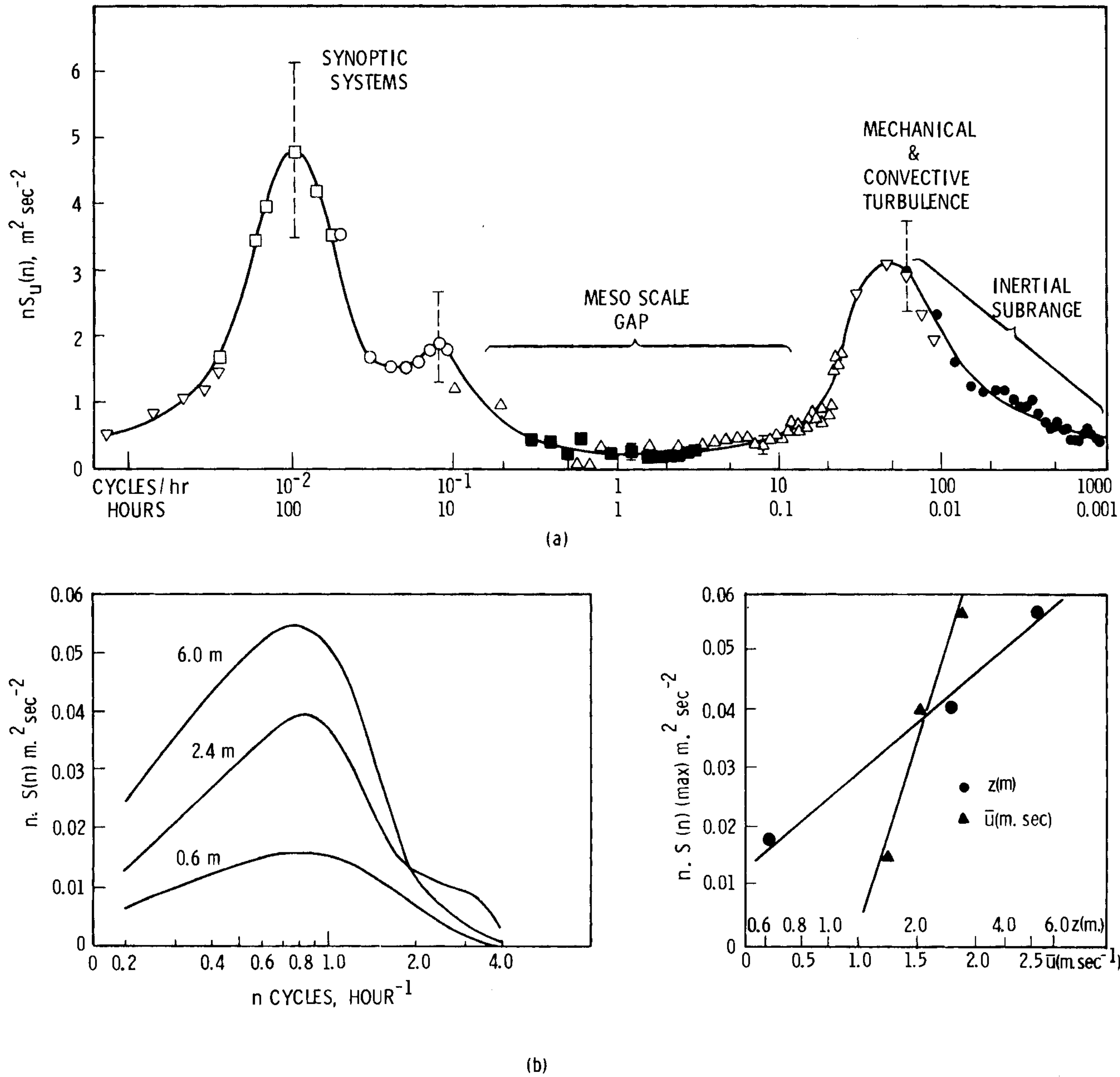


FIGURE 2.44. a) Schematic Spectrum of Wind Speed Near the Ground Illustrating the Mesoscale Spectral Gap, b) Logarithmic Energy Spectra and Variation of Maximum Spectral Density with Height and Mean Wind Speed in the Mountain Wind over Pietermaritzburg on August 16-17, 1961 (adapted from VanderHoven 1957 and Tyson 1968)

According to Tyson, the mountain wind represents a topographically induced thermodynamic wind of purely local origins that is able to support eddy energy in the range of frequencies between convection and mechanical turbulence on the one hand, and diurnal and large scale circulation on the other. Short period fluctuations are damped in the absolute stability of the mountain wind and surges with periods the order of 1 hour contribute maximum turbulent energy in valleys

2.8.4 Heat, Momentum and Moisture Fluxes

Heat (Q_H), moisture (Q_E) and momentum fluxes are the vertical and horizontal transport by the atmospheric turbulence field of sensible heat, moisture and momentum in the surface or planetary boundary layer. Munn (1966) has reviewed the present theories of these various fluxes as they pertain to homogeneous surfaces.

One example of a momentum flux is the momentum bursting process. This process is an event in which a viscous instability first occurs in the sublayer region of the boundary layer and eventually moves away from the surface. Because of these sudden concentrated bursts of momentum, gravity waves in the atmosphere can be quickly converted to highly turbulent flow extending to several hundred meters. The momentum burst implies a burst of turbulent shear stress, but heat, moisture, and other properties may also be transported (Mollo-Christensen, 1973).

Bursting phenomena have been observed in the laboratory as a series of rapid momentum bursts appearing in the form of horseshoe vortices that are advected downstream. Offen and Kline (1974); Corino and Bradkey (1969); Kim, Kline, and Reynold (1971); and Rao, Marasimha, and Narayanon (1971); and others have observed and measured bursting rates in the laboratory. Observations of momentum bursting in the atmospheric boundary layer have been reported by Businger (1972), Schubert (1975), Dorman (1971), and Schubert and Pepper (1976). Schubert and Pepper's study used real time measurements with an acoustic sounder to investigate the conditions under which atmospheric momentum bursting occurs. Results indicated abrupt transition from a laminar to a turbulent boundary layer as well as gravity-wave generation caused by shear instability (see Section 2.7.4 also).

Information on momentum and other fluxes in complex landforms is not as available. Wooldridge and Orgill (1978), and Wooldridge et al. (1976) have obtained limited measurements on momentum fluxes within selected valleys of Central Colorado. Figure 2.45 shows the momentum flux profiles for different flow conditions for the Gore and Eagle River Valleys. Rapid transverse and vertical mixing can occur near the ridge tops when upper-level flow blows with a moderate component across the valley. At times the airflow in the valley exports momentum to the flow aloft. A general separation of the air from ridge tops is suggested as a possible mechanism for the upward momentum export. The results also show that momentum can be added to the valley flow system. This import was generally contained in shallow layers well above midslope.

These tentative results show that the vertical turbulent flux convergence of momentum downward into valley air from airflow aloft is important in determining the resultant valley flow. When the surface radiative heat exchange is weak, this flux may strongly influence the valley flow and if the radiative heat exchange is strong, these fluxes may still have a modifying effect on the resultant valley airflow.

2.8.5 Free and Forced Convection

Generally there are three types of convection: forced, free, and natural. Forced convection occurs when air blows across a rough surface causing turbulence to smooth away large temperature differences. As the wind increases or as the ground becomes rougher, forced convection becomes more vigorous. Free convection is caused by density or buoyancy differences within a moving fluid. It develops as a result of uneven heating of the ground. Natural convection occurs over a heated surface when there is no wind.

Free and forced convection are seldom separate mechanisms but often interact with each other such as in the case of an eroding temperature inversion or a developing cumulus cloud over a mountain. In this latter case, Braham and Draginis (1960) and Silverman (1960) attempted to disentangle the "orographic

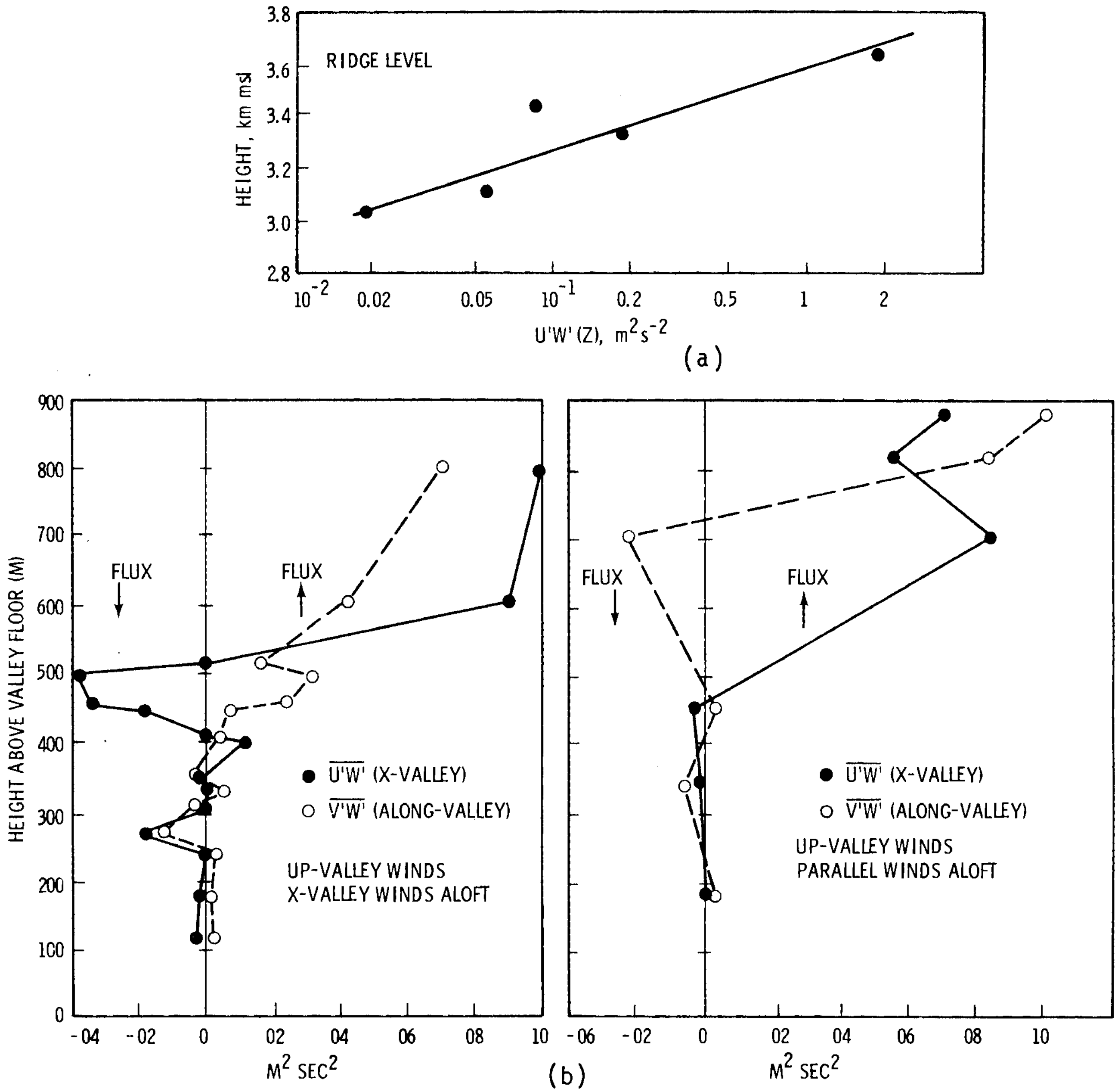


FIGURE 2.45. Vertical Dependence of Momentum Flux Within a) the Eagle River Valley and b) Gore River Valley in Colorado as Computed from Tracks of Superpressured Balloons (from Wooldridge and Orgill 1978, and Wooldridge et al. 1976)

barrier" (forced convection) and "high-level heat source" (free convection) effects as they may combine in the development of summer convective clouds over mountains. A series of measurements were made over the Santa Catalina Mountains of southern Arizona. Data obtained at sunrise show very clearly the barrier effect that forces air to ascend about 300 m in crossing the mountain crest. After the mountain slopes become heated by insolation, a convection core forms over and slightly downwind from the ridge. This core serves as the root of several cumulus clouds. The convection column had its origin in the valley-slope breeze that converges into a massive column upon reaching the mountain top. This column continues as free convection for a short distance above the top of the mountain. Under favorable conditions of moisture and temperature, the free convection penetrates to greater heights allowing the clouds to develop vigorously and to become larger.

A numerical model for investigating the effects of ambient winds on the initiation and development of cumulus clouds over an idealized mountain ridge has been developed by Orville (1968). Preliminary conclusions were that cloud initiation is stimulated by the ambient winds with cloud development initiating earlier and forming over the downwind slope rather than directly over the peak. However, development is slower than for the case of no initial wind. Also, a case is presented in which the heating of the upwind slope creates a wave in the airflow upwind of the mountain ridge that generates clouds that advect downstream.

2.8.6 Summary

At the present time there have only been relatively sparse measurements of wind spectra, turbulent intensity, and various turbulent fluxes in complex landforms. The interpretation of these data in terms of real physical quantities is not rigorously established. Turbulence measurements are difficult to obtain in complex terrain settings but the establishment of a good data base is a necessity in order to improve understanding of mechanical and convective turbulence transport processes.

The modeling of heat, momentum, and moisture fluxes over relatively uniform terrain is quite extensive but such modeling for complex terrain is rather limited. An important aspect to this type of modeling will be the consideration of forest canopy flows and their exchange processes. This is discussed briefly in the next section and will complete Section 2.0.

2.9 FOREST CANOPY FLOW AND EXCHANGE PROCESSES

Wooded areas or forests are active meteorological regions. A forest can influence the surface boundary layer and tower layer where it interacts with the local or lower-level boundary layer winds. Within or near forests the exchange processes can be vigorous despite light winds. In addition, the forest canopy can be penetrated by heat, water vapor, and momentum. Moisture conditions and transpiration rates may differ from other areas because of the type and extent of foliage and undercover. In hilly or mountainous country there are also differences in slope and aspect. In short, a forest contains important sources and sinks of heat, moisture and momentum that depend to a large extent on physiological factors.

A standard reference on forest meteorology and climatology is Geiger (1966) who presents some informative data on terrain effects. Munn (1966) and Glesinger (1962) also provide useful information. In the past, studies have generally considered the forest environment in terms of the uniform forest, the forest edges, and clearings. In the following, temperature, winds and turbulence in forests are discussed in these categories; a distinction is also made between studies over semi-level ground and those in complex landforms.

2.9.1 Uniform Forest Stand

Recent results of the thermal structure within and above a forest canopy appear to confirm the general aspects of earlier studies (Hosker et al. 1974). Temperature measurements were made within a dense stand of loblolly pines with a mean tree height of about 17 m. On a clear day a very unstable temperature gradient occurs above the trees, while a strong inversion (~8°C) develops below the crowns. The position of the daytime temperature maximum in the tree-tops responds to the solar elevation, eventually descending about 2 m to the region of maximum foliage density as the sun's rays penetrate deeper into the tree crowns. At night, the sub-crown region becomes weakly unstable, but the atmospheric layer above the trees is then stable. On an overcast and rainy day, the strength of the temperature inversion beneath the tree crowns is less that 1°C. Breaks in the overcast result in rapid above-canopy and upper-crown heating. A temperature maximum is quickly established in the central portion of the crown, with strong stable gradients in the lower crown.

In a dense forest, neither sunshine nor wind can penetrate to a significant extent, so the air in the trunk area remains cool and moist, contrasting with the warmth in the zone of the tree tops. In a thin forest, the sun's rays at noon are able to penetrate farther into the forest, and more vigorous mixing brings the heated air aloft down more quickly. Conversely, at night, strong temperature inversions develop in a thin stand similar to bare ground (Geiger 1966).

Vertical profiles of air temperature were obtained in a forest canopy located on a mountain slope during the night with little or no cloud cover (Bergen 1971 and 1972). Temperature profiles showed a local maximum, well within the forest canopy. Behavior of this maximum indicated that the warming is due to subsidence heating caused by topographically induced divergence of the density flow in the canopy. Earlier measurements of a similar nature were reported by Baumgartner and Hoffman (1957), and Sauberer and Trapp (1941).

Trees affect the wind in at least three ways. Airflow around individual trees and their component parts creates wakes or separation zones that exist downwind (Meroney 1968). Airflow over dense stands of trees creates internal momentum and thermal boundary layers and also exert a braking action on the lower level winds. A number of studies on winds within and over forests have generally shown typical wind profiles such as those shown in Figures 2.46 and 2.47 (Reifsynder 1955; Geiger 1966; Meroney 1968; Tourin and Shen 1969; Oliver 1971; Martin 1971; Bergen 1971, 1971/1972, and 1969; Raynor 1971, and Smith, Carson, and Oliver 1972). A special feature, which may often be observed in forests, is the existence of air movement near the trunk particularly when the wind can blow in through the open borders of the stand or when the winds are strong. The extent to which wind can penetrate a forest depends

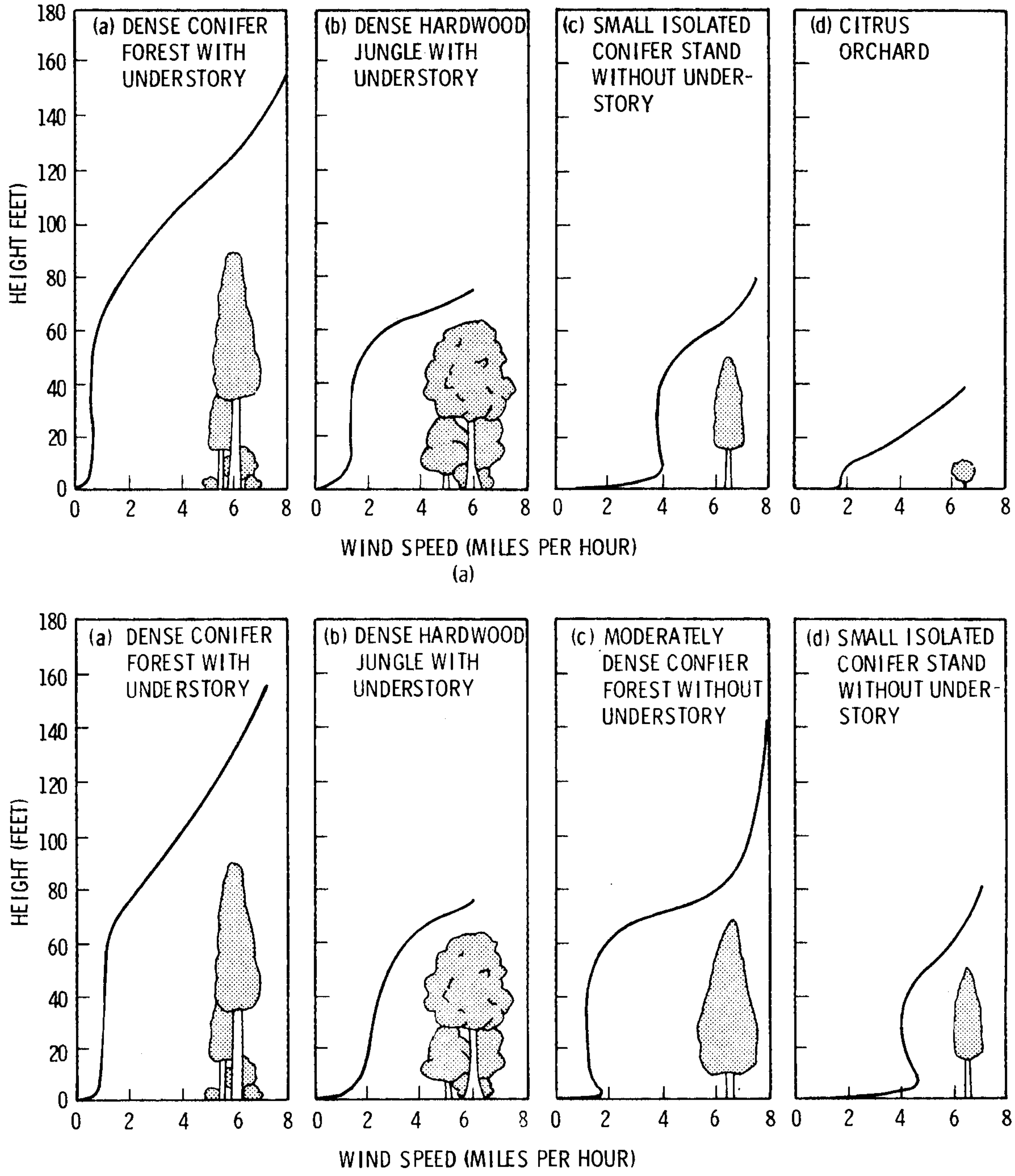


FIGURE 2.46. Wind Speed Profiles Over Different Forest Canopies: a) Comparative Wind Speed Profiles During Temperature Inversion conditions and b) Comparative Wind Speed Profiles During Lapse (neutral) conditions (adapted from Reifsnyder 1955)

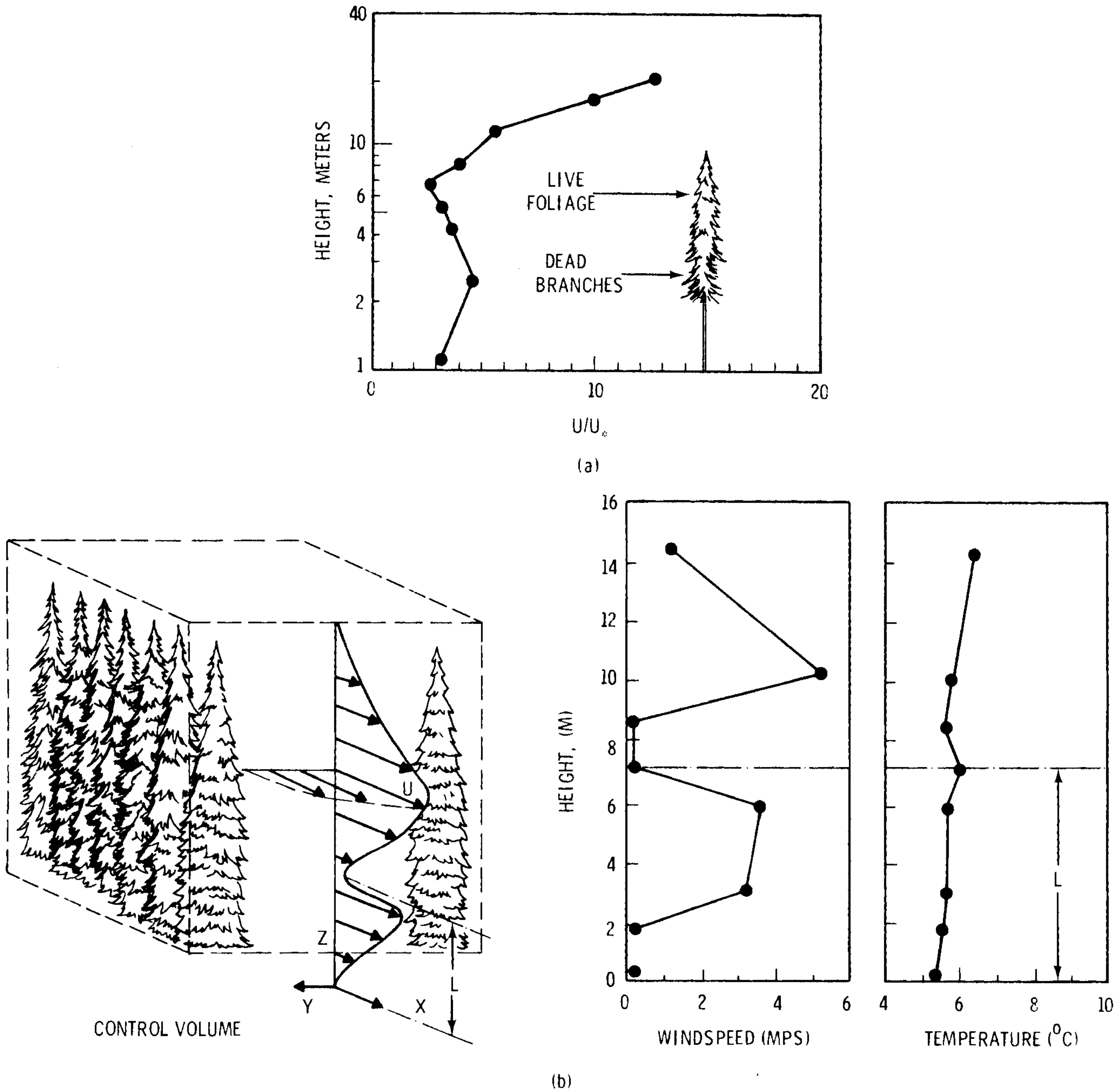


FIGURE 2.47. Wind Speed Profiles Over Forest Canopies in Mountainous Areas: a) Wind Speed Profiles in a Lodgepole Pine Stand at Fox Park, Wyoming and b) Average Vertical Profiles of Wind Speed and Temperature at Fraser, Colorado (adapted from Bergen 1969 and 1971)

on several factors. Thickets and saplings bring the air to a complete halt. In old stands, the type of wood and the degree to which the crown is closed are of importance.

Martin (1971) considered the effects of topography and atmospheric stability on wind profiles over a forested area. At night, when the atmosphere is stable, variations in profile shape above the forest are associated largely with site properties. During the day the variations are less, indicating that convective turbulence tends to control the profile shapes and to mask the effect of site irregularities. Momentum transfer through the canopy reaches a maximum at noon. At night the momentum transfer drops to 75% of the daytime maximum value and displays a marked variation with direction.

Turbulence measurements within and over forest canopies have been made by Allen (1968), McBean (1968); Tourin and Shen (1969); Bowne, Entrekin, and Anderson (1970); Thompson (1979); and Raynor, Hayes, and Ogden (1974). Meroney (1968) and Sadeh and Kawatani (1978) used model forest canopies to simulate the meteorological characteristics of typical forests. Meroney made measurements on longitudinal turbulent intensity profiles over the canopy and some of these results were comparable with Tourin and Shen's field measurements.

Turbulence measurements by Raynor, Hayes, and Odgen (1974) showed that the intensity of turbulence within a forest reaches a maximum at midcanopy level. At the midcanopy level, the vertical component equals the lateral and is about three times as great as in the free air above the forest and eight times as great as in the trunk space. At the canopy level, mechanical turbulence is enhanced by protruding tree tops of different heights and densities. Downward motions are noticeable in the wake of these obstacles. Convective turbulence is generated by unequal heating between the sunlit and shaded sides of each tree and by holes and openings in the canopy. These combine to convert much of the horizontal motion above the forest into vertical motion that penetrates the canopy.

McBean (1968) measured the vertical component of turbulence in a forest canopy located in a small watershed surrounded by mountains. His results indicated that the vertical component of turbulence is as high or higher in the forest than over open ground. Magnitudes of temperature and humidity fluctuations were also higher and were attributed to the inhomogeneity of the forest. An energy balance was not obtained and it was suggested that in case of inhomogeneous terrain that spatial as well as time averages of the turbulent heat fluxes and the net radiation are necessary for obtaining a good energy balance. Energy or radiation budget studies have been conducted by Federer (1968), Stewart and Thom (1973), Thom et al. (1975), and Thompson (1979).

2.9.2 Forest Edges

Depending on the orientation of the forest edge, the ground and the air between trees may be sunny or shady, windy or calm, warm or cold, moist or dry. According to Geiger (1966), the forest edge climate develops from two sources. First, it is a transition climate, and second, the edge of a forest forms a high step in the topography, which takes up or wards off sunshine, wind, and rain.

In relation to wind influences at forest edges, Geiger (1966) and Pfeiffer (1938) distinguish between passive and active influences of a forest on the wind. The passive forest is an obstacle to the airflow. On the windward edge of a forest stand the airflow can be stagnated or eddies may develop. With an open stand of trees, the wind can penetrate into it in gusts but is quickly slowed down by the tree trunks or undergrowth. Streamlines are cramped together in the border zone above the forest. On the leeside of a forest stand there can be an area sheltered from the wind (Bergen 1979). When the wind is blowing at an angle to the stand edge, it acts as a steering line, and wind increases in the outer edge area.

The active influence of a forest is illustrated when the temperature differences built up between it and the open country produces an airflow tending to equalize this temperature contrast. During the day, the air near open ground is heated while the air under the crown of the forest remains cool. Thus, cool air may flow out from the trunk area as a daytime forest breeze. In contrast, a field breeze at night has not been observed. The strong braking effect of the trees prevents its development. In mountainous wooded country, a night "forest wind" occurs in the open country, representing the downflow of cooler air formed over the radiating crown area. However, this is a component of the normal slope wind caused by nighttime cooling as discussed in Section 2.4.1.2.

2.9.3 Forest Clearings

To rejuvenate the forest, foresters make improvement cuttings, thinnings, or burnings to clear the way for new growth. In addition, forest fires may reduce large areas of forest to relatively sparse regions of vegetation. Naturally, such clearings modify a forest's micrometeorology depending on the size of the clearing. The size is measured according to the ratio of the diameter (D) of the clearing (assumed circular) to the mean height (H) of the surrounding stand. The ratio D:H is called the clearing size index. Geiger (1966) indicates that the microclimate of a clearing may change appreciably when the clearing size index exceeds 1.5.

The wind speed distribution in and near an isolated narrow forest clearing was measured and modeled by Bergen (1976). The clearing was approximately a rectangle 10-m by 50-m cut in a 10-m high lodgepole pine stand. Prior to making detailed wind measurements, cinematic observations were made of the behavior of multiple smoke plumes in the clearing. The cinematic results indicated a continuous alternating between separated and unseparated flow. The frequency of the alternating flows agreed fairly well with the eddy shedding frequency of the lee canopy region of the clearing, if regarded as a flat plate in uniform flow. The flow sequence included a central vortex that closely resembled in form that found in square notches but with higher angular velocities relative to the upwind friction velocity. This vortex appeared to dominate the distribution and direction of the maximum speeds and surface shear stress in the clearing.

Wind measurements were made when the wind direction above the canopy was perpendicular to the long axis of the clearing and were used in a continuity calculation to establish the velocity field for the flow in and downwind of the clearing. Minimum speeds occurred at the clearing center and at the midcrown region on the lee edge of the clearing. Speed maxima occur at subcanopy levels on either edge and above the lee edge. Clearing effects extend behind the clearing to at least 25 m, but are only slightly apparent upwind of the clearing (Figure 2.48a). The continuity calculation indicated a separation of the flow beginning at the midclearing flow and extending to the upper surface of the lee canopy. Reattachment occurs 4 or 5 m behind the lee clearing edge, with strong reverse flow apparent upwind from that point (Figure 2.48b). Bergen compared his results with earlier studies of Pfeiffer (1938), Krecmer (1967), Fojt (1964), and Geiger (1934, 1966) but found that the clearing shapes, sizes, and stand heights varied so much that comparisons were difficult. Nevertheless, some similarity in the rotor flow and throughflow states were noted.

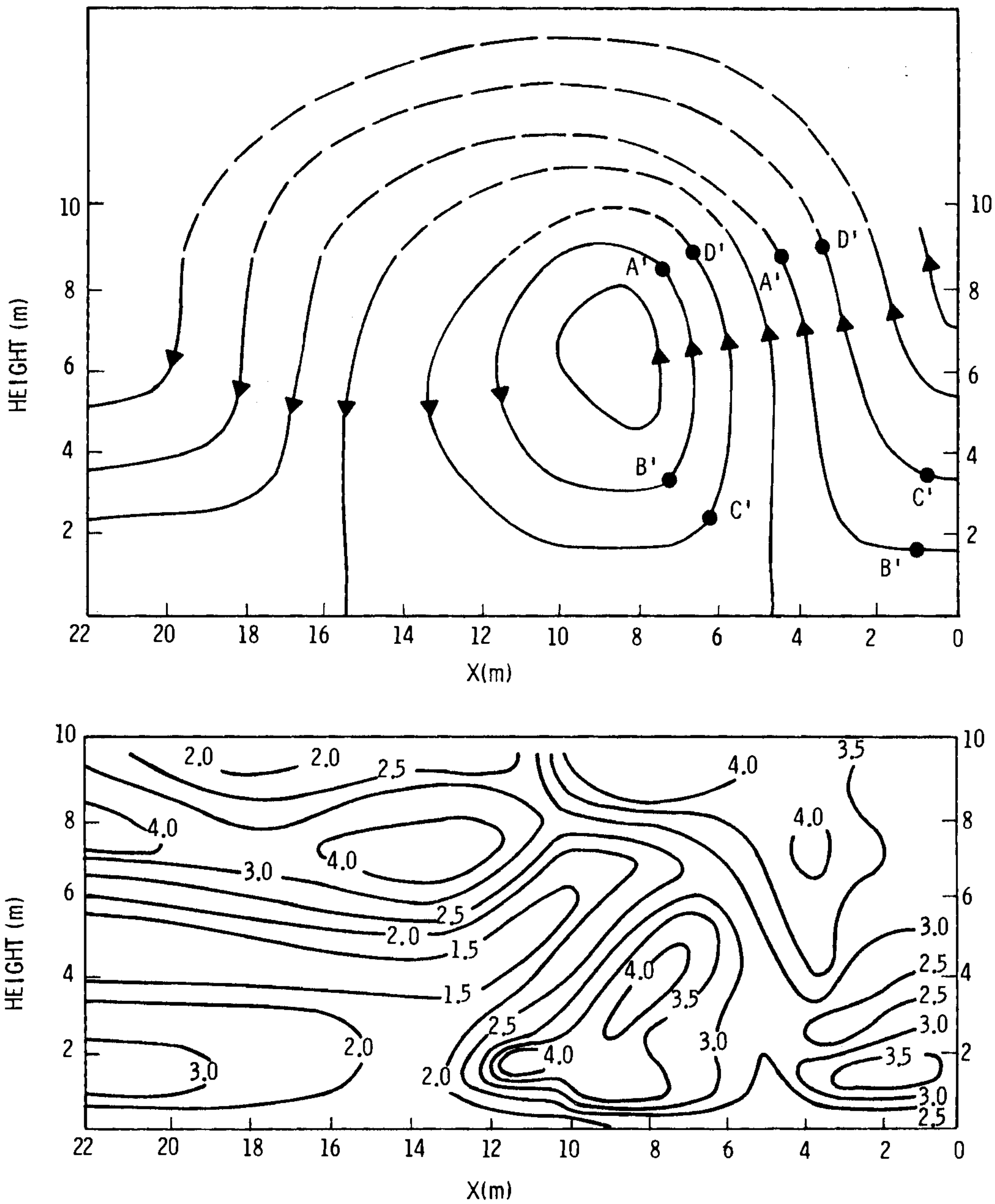


FIGURE 2.48. a) Estimated Streamline Pattern for the Flow Across a Rectangular Forest Clearing and b) the Scaled Speed Distribution Across the Clearing (X is the distance from windward edge) (adapted from Bergen 1976)

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3.0 PLUME OR POLLUTANT TRANSPORT AND DIFFUSION IN COMPLEX LANDFORMS

Turbulent diffusion[a] in the atmosphere is a complex phenomenon caused by various sized eddies in the atmosphere (see Section 4.2). Those eddies with larger dimensions than a pollutant plume will tend to transport the plume while those near the same size or smaller than the plume will tend to spread or diffuse the plume. Figure 3.1a shows a plume being released into the atmosphere. As the plume is released, two parcels of air (e.g., constant-volume balloons) are released into the plume. The mean wind velocity affecting the plume is represented by $\overline{V}$ and ΔX, ΔY, ΔZ are the longitudinal, lateral, and vertical separation distance of the two ballons, respectively. In Figure 3.1b, the transport of the plume and balloons is represented by the travel time or distance as calculated from the mean wind velocity $\overline{V}$. The spreading of the plume or dispersion is represented by the total separation distance of the plume or balloons; i.e.,

$$R = \left[(\Delta X)^2 + (\Delta Y)^2 + (\Delta Z)^2\right]^{1/2} \qquad (3.1)$$

This equation represents the result of turbulent eddies and wind shear. However, such a decomposition of the flowfields into mean transport and turbulence diffusion is rather artifical and is used mainly for convenience.

In discussing atmospheric diffusion, a distinction is often made between two systems of reference in respect to averaging times: the spatially fixed or Eulerian system and the particle-attached or Lagrangian system (Gifford Chapter 3 AEC 1968). The Eulerian fixed-point system is commonly used by field experimentalists in conducting measurements of plume diffusion from

(a) Some authors prefer to use disperion when refering to particle diffusion reserving diffusion to describe gas, heat, and momentum transport. Diffusion and dispersion are closely related, differing only in that for dispersion, the possibility exists of effects arising from the size and inertia of the particles. In this section, dispersion will be used when particulate plumes are suggested.

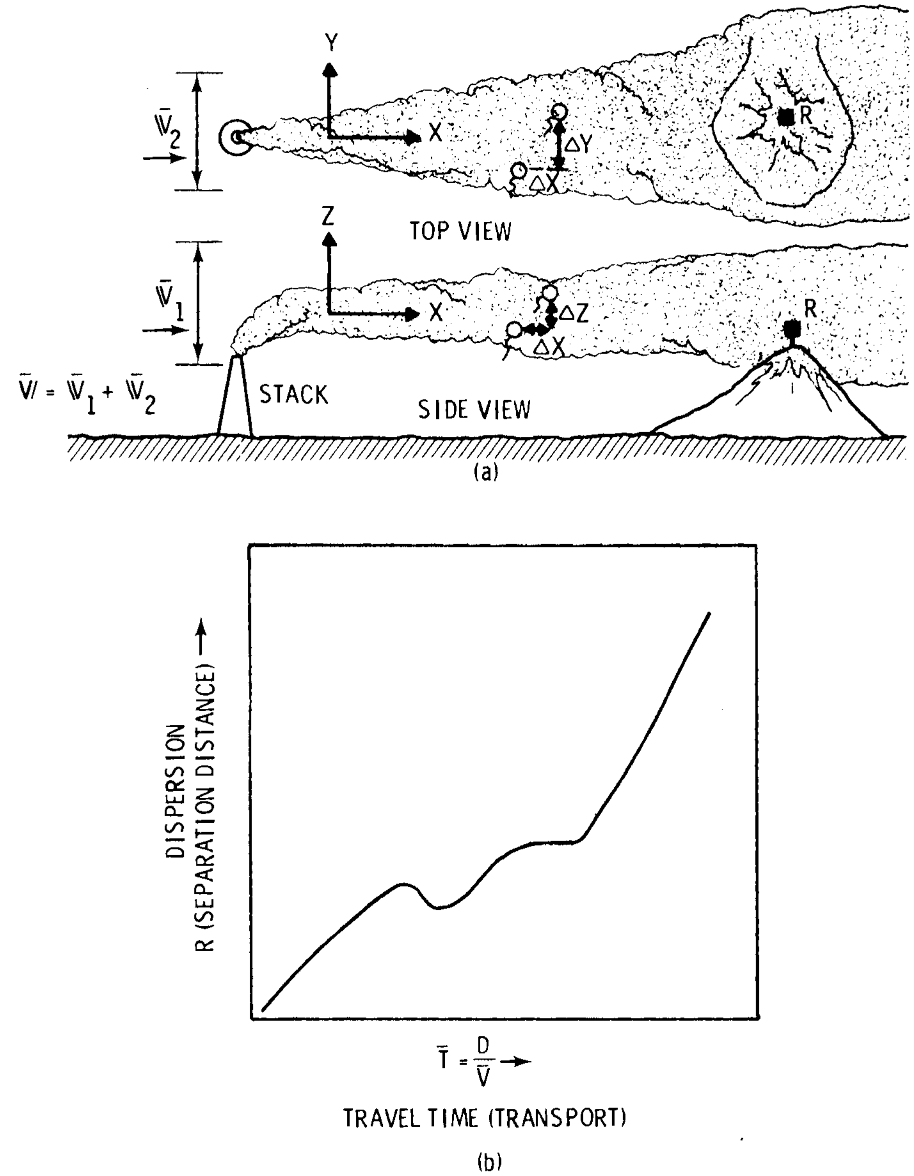


FIGURE 3.1. Hypothetical Illustration of Transport and Dispersion of an Effluent Plume: a) Top and Side Views of a Dispersing Plume. $\bar{V}$--average total wind effecting plume; Δx--longitudinal distance between balloons or tetroons; Δy--lateral distance; Δz--vertical distance; R--receptor site. b) relationship between dispersion and travel time (transport).

sources and fixed sampling locations (Receptor R in Figure 3.1a). In this Eulerian system, all scales of eddies act on the plume simultaneously and the distinction between transport and diffusion becomes a matter of travel versus sampling time. However, the Lagrangian system indicates how turbulent eddies with scales on the order of the plume produce dilution through entrainment of ambient air, while the larger eddies act to transport the plumes (see Figure 3.1).

Factors considered in any problem of diffusion meteorology include: process factors, source or stack factors, meteorological factors, topographic factors and depletion and resuspension factors. Table 3.1 lists the variables that have to be considered in relation to these factors. In the case of level or uncomplicated terrain, the topographic factors in the transport and dispersion problem are essentially nonexistent except for roughness and vegetation. In level terrain many of the effects listed in Table 3.1 have been documented in past research and reference works (Pasquill 1962, Turner 1967, Stern 1976, AEC 1958, U.S. Weather Bureau 1955, Magill et al. 1956, and Smith 1968). In complex landforms the topographic factor in diffusion problems becomes the dominant variable because it can modify all the other factors except the process factors which remain unaffected. In the following sections some of the effects of topography on source, meteorological and depletion factors in the transport and dispersion problem will be reviewed.

3.1 SOURCE FACTORS

Sources in complex landforms may not differ significantly from what exists in level terrain, i.e., point, line, area and volume. Single or multiple sources can be elevated or at ground level. The resulting plume can be either passive or buoyant. Elevated and multiple point sources such as coal-fired steam-generating electric power plants are becoming rather common in complex landforms. Line sources are not as common but lines of cooling towers or inter-state highways may approach this type of source. Area sources refer to locations such as cities, industrial centers, cooling ponds, or forest fires. In addition, terrain could produce secondary sources, i.e., area, line or volume sources, when the initial pollutants become trapped by some landform (Figure 3.2).

TABLE 3.1. List of Variables that Affect Any Problem in Transport and Dispersion

- Process Factors
  1. Emission rate
  2. Temperature of emission products
  3. Form of emission product--dust, fume, mist, spray, or gas
  4. Concentration of gaseous and particulate matter
  5. Particle size distribution and terminal velocities
  6. Agglomerating characteristics
  7. Chemical properties
  8. Toxicological or nuisance properties

- Source Factors
  1. Source type, i.e., point, line, area, volume
  2. Source height
  3. Source diameter and configuration of exit
  4. Source emission velocity

- Meteorological Factors
  1. Wind speed and direction
  2. Temperature and humidity
  3. Atmospheric stability
  4. Wind profile
  5. Turbulence
  6. Mixing height
  7. Cloudiness and preciptation
  8. Heat transfer

- Topographic Factors
  1. Relationship of source to surrounding structures and terrain
  2. Local and distant landforms
  3. Bodies of water i.e., lakes, coastal areas, and rivers
  4. Roughness (vegetation, etc.)

- Depletion and Resuspension Factors
  1. Gravitational settling
  2. Plume impaction
  3. Chemical transformation
  4. Wet and dry scavenging
  5. Electrostatic attraction
  6. Coagulation
  7. Resuspension

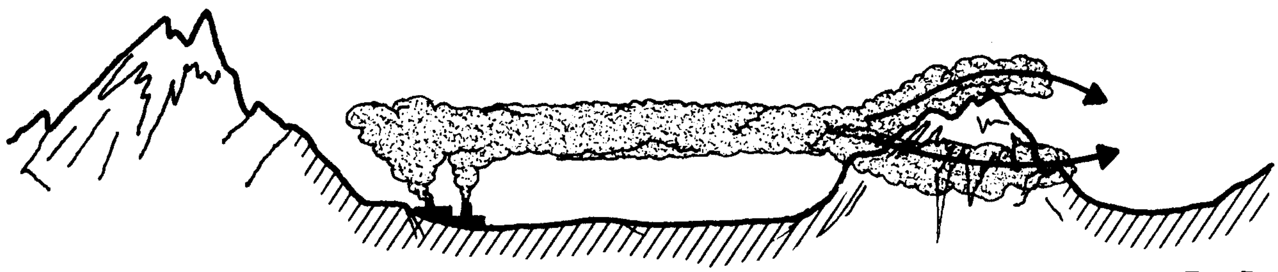


FIGURE 3.2. Illustration Indicating Secondary Source Characteristics of a Valley. Pollutants accumulate in Valley A and act as a line, area, or volume source for Valley B.

The initial plume dispersion depends on:

- the plume rise relative to the mean motion of the air as a result of the buoyancy and initial vertical momentum of the plume.
- diffusion because of turbulence in the air.
- aerodynamic effects caused by stacks, buildings, and landforms.

The height of plume rise depends on the rate of heat emission, rate of momentum emission, the wind speed at the plume level, the distance downwind of the source, the height of the source, and the variation of ambient temperature with height (see Briggs 1959 and 1971).

Stack plumes can be classified according to the initial plume dispersion as a result of turbulence and stability. One classification scheme relates the spreading of the stack plume to the vertical temperature gradient and wind velocity profile. These types include: looping, coning, fanning, fumigation, lofting, thermaling, flagging, downdrafting, puffing and bifurcating (Scorer 1978; Smith 1968). At source locations, the observation of visible plumes may provide helpful information on the effects of surrounding terrain.

In addition aerodynamic effects of local terrain features near a source can modify plume behavior by affecting the initial plume rise, wind patterns, and turbulence. These effects are difficult to quantify because of the large variety of possibilities. However, some attempts have been made to develop a set of mathematical statements. Stumke (1964a and b) has provided a method for correcting effective stack height for a simple step or cliff in terrain, but only for streamline flow. Another technique for assessing terrain and building effects near sources uses wind tunnels and stratified water channels (AEC 1968; Lin, Liu, and Pao 1974; Liu and Lin 1975 and 1976; Cermak et al. 1978). This technique usually provides semiquantitative results on the effects of terrain on plume rise and initial plume dilution.

In some cases, illustrative examples based on observations are about the only information available on possible effects of a particular source, landform type or meteorological conditions. Hewson, Bierly, and Gill (1961) has shown the expected behavior of a stack plume near a steep cliff or bluff when the prevailing flow is toward the cliff under stable and unstable stability conditions (Figure 3.3a). When the prevailing airflow is away from the cliff or bluff, then aerodynamic influences may affect a stack plume as shown in Figure 3.3b. If the source is close to the bluff or cliff, then aerodynamic eddies may bring the plume down to groundlevel in a short distance. However, if the source is somewhat distant from the cliff and the prevailing airflow over the plateau is slightly stable or neutral, the plume may be carried to the ground in a downwash because of a separation eddy in the lee of the cliff (Figure 3.4).

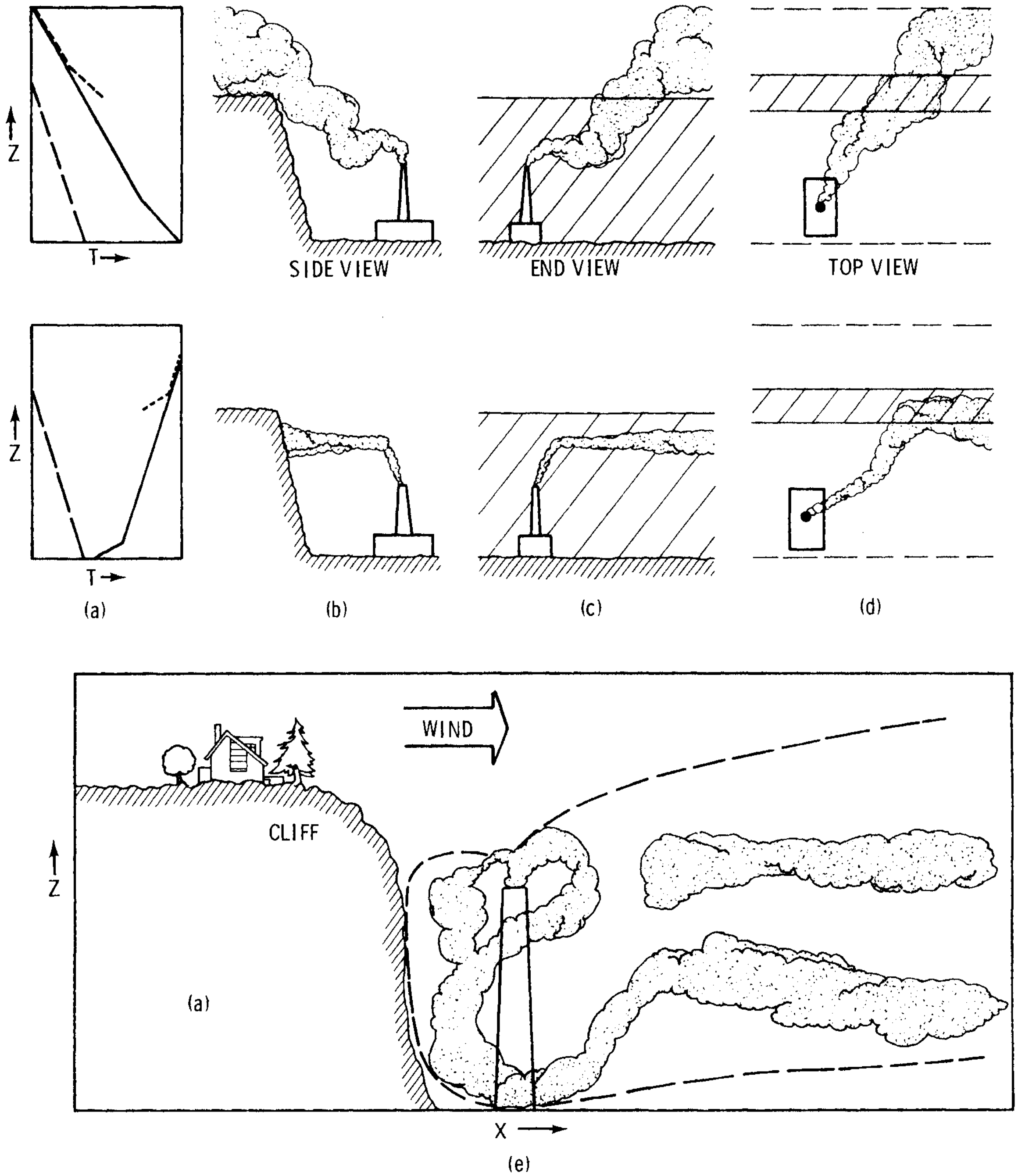


FIGURE 3.3. Aerodynamic Effects of Terrain on Plume Rise and Initial Transport and Diffusion. a,b,c,d) Plume behavior near a steep bluff when the air is unstable (above) and when it is very stable (below) and e) when the prevailing airflow is away from the bluff and the air is neutral or slightly stable (adapted from Hewson et al. 1961)

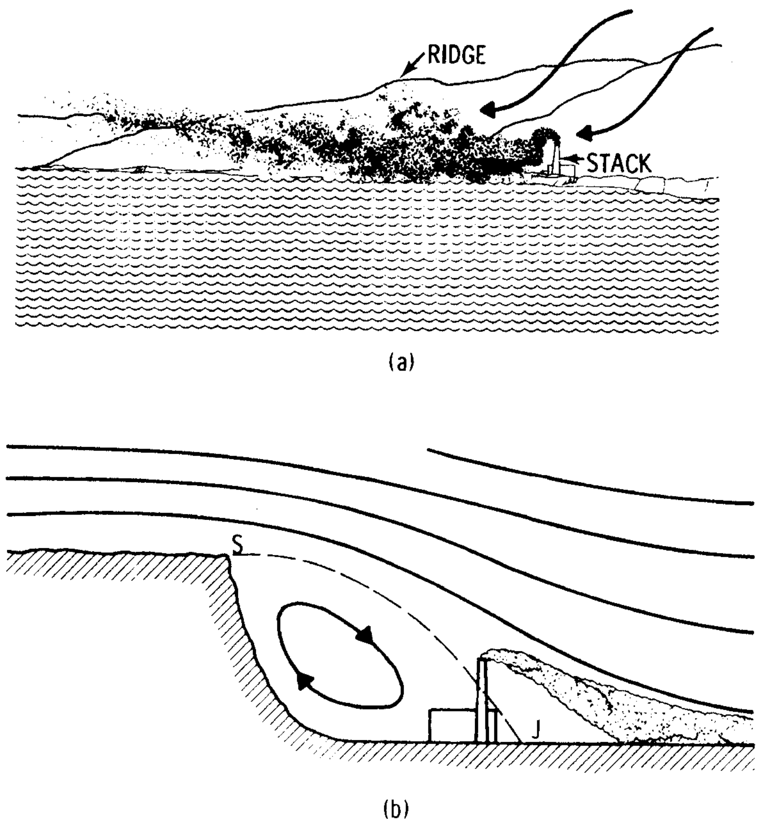


FIGURE 3.4. Downwash of a Plume as the Result of of Airflow Separation Near a Cliff or Ridge: a) a Plume Carried Immediately to the Water Surface in the Lee of a Wooded Ridge and b) a Schematic Diagram Showing the Effects of a Separation Eddy on a Plume Just Lee of a Cliff (adapted from Scorer 1972)

3.2 TOPOGRAPHIC AND METEOROLOGICAL FACTORS

One of the major problems of air pollution monitoring and modeling in complex terrain is understanding and simulating the plume behavior because of topographic and meteorological factors. In order to understand or simulate plume behavior in complex landforms it is important to identify and understand the physical role that different terrain-induced airflow phenomena plays in the transport and dispersion process. Section 2.0 identified and discussed the physical aspects of several of these phenomena. In the following section such phenomena as orographic-dynamic effects, local winds and turbulence (dispersion) will be reviewed in respect to how they can influence transport and dispersion in complex landforms. In addition, terrain effects will be briefly assessed for dispersion over forest canopies and in long-range transport although the data base is limited.

3.2.1 Orographic-Dynamic Effects

Orographic-dynamic effects will be defined as any terrain-induced phenomena which affects airflow and pollutant patterns caused by the physical presence of particular landforms. These effects include such phenomena as streamline or plume deformation, channeling, impaction, stagnation, trapping, and downwash.

3.2.1.1 Streamline (Plume) Deformation

As discussed in Section 2.1.1, the airflow streamlines in and over complex landforms undergo deformation as a result of topographic obstacles. (See Figure 3.5 for a few simple illustrations.) Figure 3.5 shows the deformation of a hypothetical plume upstream from a cylinder and sphere based on potential flow theory. According to Egan (1975) plume centerlines originating far below the crest height of the obstacle have a height of closest approach about one half of the initial plume height above the surface. The increase of velocity over the crest, however, results in a nearly 2:1 decrease of the vertical spacing of streamlines. Thus, a plume following the trajectory would be close to the surface at the crest, but in the absence of diffusion, narrower in the vertical dimension.

Hunt, Puttock, and Snyder (1979) and Hunt and Mulhearn (1973) have also considered the effects of two- and three-dimensional obstacles on transport and turbulent dispersion of plumes. Their results show that because streamlines approach the surface of a three-dimensional hill much more closely than the surface of a two-dimensional hill, the maximum surface concentration on the hill can become greater than in the absence of the hill. However, this only occurs for a limited range of source heights (Figure 3.5b).

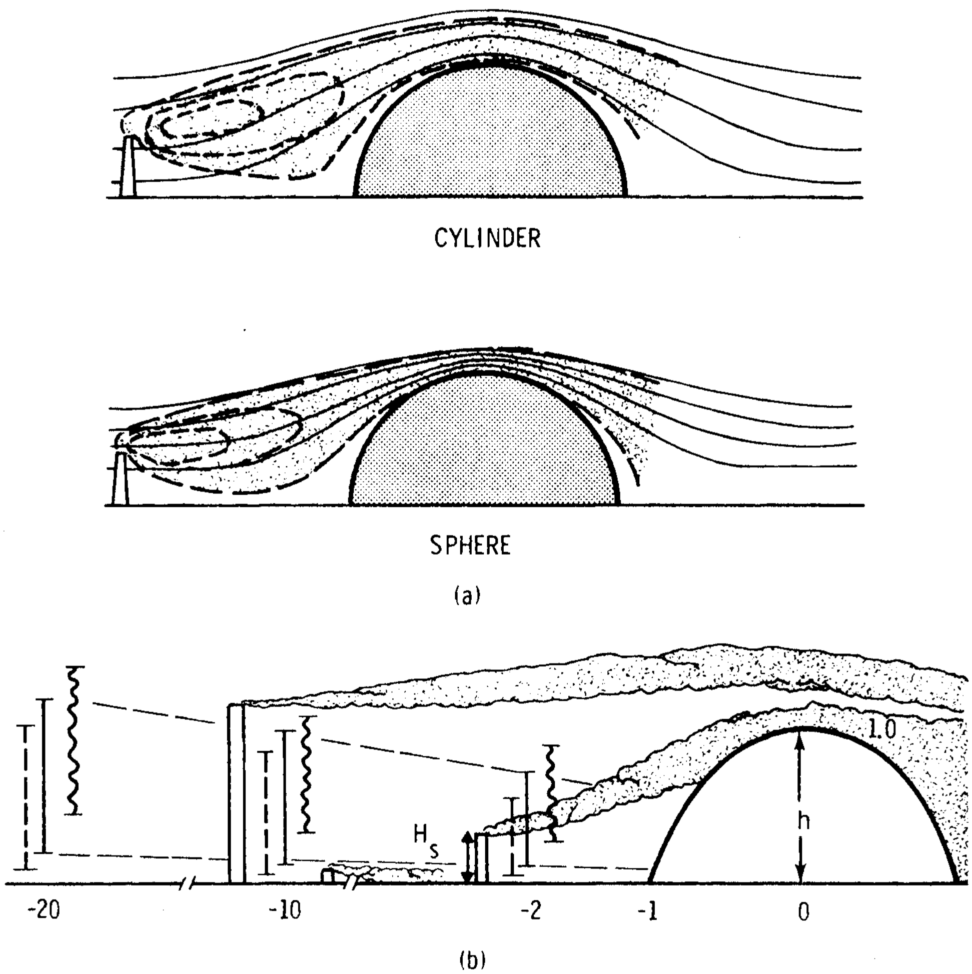


FIGURE 3.5. Examples of Plume Deformation as a Result of Simple Topographic Obstacles: a) Plume Deformation Caused by Potential Flow Over a Cylinder and Sphere and b) Potential Flow Over a Hemisphere with constant Diffusivity. Demonstration of the "window" or range of values of H_s,h for which the maximum ground-level concentration is increased by a factor of 2. If H_s is too large the plume misses the hill; if too small the plume touches the ground upstream (adapted from Hunt et al. 1979)

Deformation of plumes result from the spatial and temporal variations of wind speed and direction: Wind speed variations with height stretch fluid elements out in the flow direction and thus allow gradient transfer processes to act more quickly, effectively increasing the turbulent diffusion rates. This effect is most pronounced for instantaneous (puff) releases of pollutants and less pronounced for plumes resulting from steady release rates. Wind direction changes with height have an even larger effect on diffusion rates and act significantly on both instantaneous and steady release rate plumes. Terrain effectively increases the magnitude of the spatial and temporal variations of wind speed and direction (Benjamin et al. 1977).

3.2.1.2 Channeling

Channeling is one of the major effects that complex landforms have on wind and therefore pollutant plumes. As stated in Section 2.1.3 channeling of plumes may occur through mountain passes, down or up canyons or valleys and along ridges. Numerous examples of channeling or inferences to channeling can be found in the literature. A good example is shown in Figure 3.6 based on Reid's (1979) data from aircraft sampling in the Eagle River Valley of Central Colorado.[a]

3.2.1.3 Impaction

In a study on the dispersion of SO_2 in the Columbia River Valley near Trail, British Columbia, Canada Hewson and Gill (1944) noted relatively high concentrations of SO_2 on valley walls, indicating the possible impaction of elevated plumes on high terrain (Figure 3.7). Recent studies by Start, Ricks, and Dickson (1974); Start, Ricks, and Wendell (1975); Rockwell International (1975); Williams and Cudney (1976); Roffman et al. (1976); and Wilson, Fabrick, and Sklarew (1977) suggests that relative high concentrations of pollutants can occur on high terrain during stable or neutral flow conditions indicating that plumes may impact on terrain plateaus such as canyon walls and cliffs. Start,

(a) Laboratory fluid or wind tunnel airflow models also show channeling effects caused by topography (see Orgill et al. 1971a, b and Liu and Lin 1976).

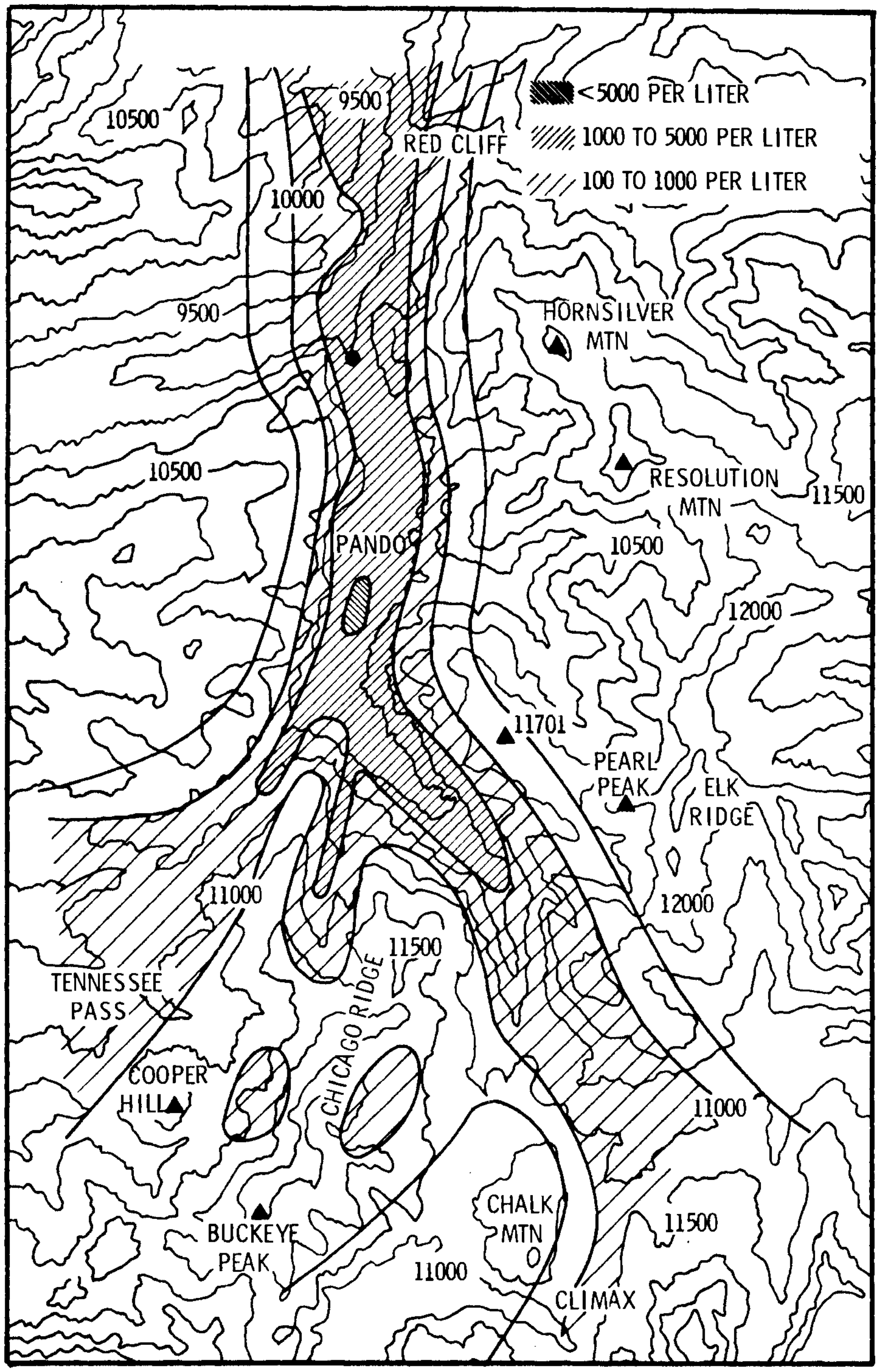


FIGURE 3.6. Channeling of Artificial Ice Nuclei in the Eagle River valley of Central Colorado, December 15, 1973 (adapted from Reid 1979).

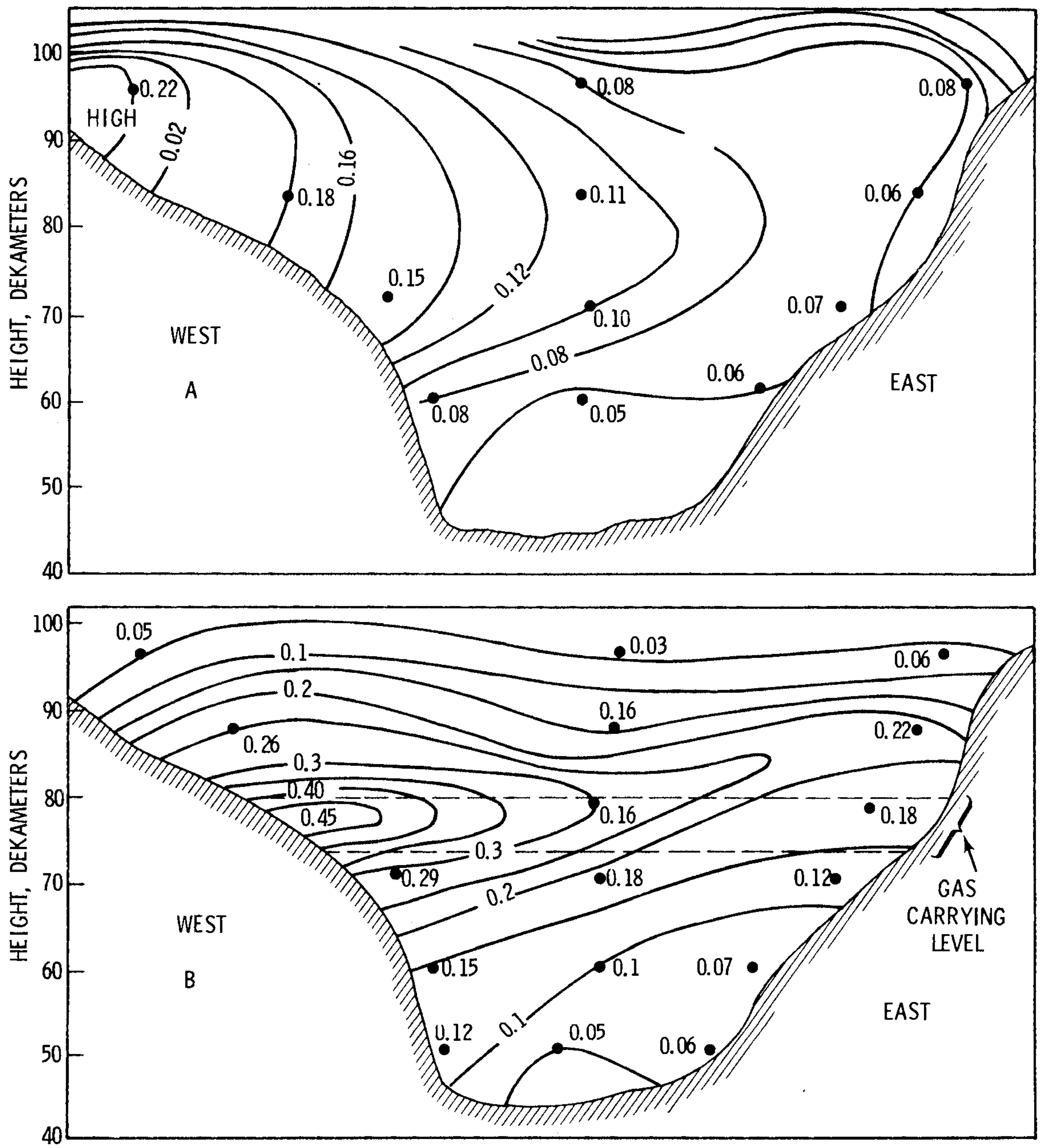


FIGURE 3.7. Suspected Impaction of SO_2 (Sulfur Dioxide) on Valley Walls in the Columbia River valley: a) 1500 pm August 1938, b) 0730 AM, February 21 to March 1939 (adapted from Hewson and Gill 1944)

Ricks, and Dickson (1974); and Start, Ricks, and Wendell (1975) feel that plume impaction does occur on terrain but the concept should be made with qualifications. A coherent narrow or filament-like plume is not likely within a steep-walled, deep canyon such as the one studied by Start et al. (1975). Plume segments from elevated sources dispersed laterally and when they approach elevated terrain can produce transient effluent concentration levels. Williams and Cudney (1976) and Roffman et al. (1976) suggest that by means of aircraft sampling elevated plumes could have impacted on terrain features varying from 6 to 55 km from the source but additional verification from ground sampling was not available. Elliot, Hovind, and Edelstein (1977); and Wilson, Fabrick, and Sklarew (1977) have developed and tested a modified Gaussian diffusion model and a numerical model (DEPICT) for examining impaction of plumes on terrain. Preliminary results indicate that such models give an improved plume dispersion assessment around elevated terrain but verification is lacking because of the small field data base.

3.2.1.4 Stagnation

As discussed in Section 2.5, stagnation conditions are especially favorable in valleys or basins. Thus, valleys or basins are poor locations for large industrial or power-generating plants if air pollution is the dominant consideration. The air pollution episodes that occurred in Donora, Pennsylvania, in 1948 and the Meuse River valley, Belgium, in 1930 resulted from inadequate ventilation of the valley because of terrain features (Stern 1968). In both cases, the valley sides were only a few hundred meters in height, but this proved sufficient to restrain the natural ventilation under the existing prolonged 4- to 5-day inversion periods. The situation could have been more critical if heavy emmissions were present in a relatively narrow valley or basin with mountains rising steeply on either side.

Meteorological conditions conducive to these episodes of prolonged periods of poor ventilation are fairly common in the mountainous western and southeastern United States (Korshover 1960 and Holzworth 1972). The conditions are associated with the stagnation of large high-pressure systems over these regions. Typically, these systems are characterized by low wind speeds, clear or foggy skies, subsidence inversions, and nocturnally produced ground-level inversions.

Wilson et al. (1976) and Leahey and Halitsky (1973) examined the problems of diffusion under low ($<2ms^{-1}$) windspeed conditions. Wilson et al. conducted a series of controlled gaseous tracer releases in heavily forested rolling terrain near Oak Ridge, Tennessee, during neutral and moderately stable conditions. In all instances, the observed axial concentrations were less than those predicted by Pasquill-Gifford curves for the appropriate stability classes. The standard deviation of lateral plume spread (σ_y) averaged six times greater than predicted values. This increase was attributed mainly to wind meander The derived vertical plume spread statistic (σ_z) averaged 5.7 times greater than the Pasquill-Gifford predicted values. Overall average dilution of ground-level peak concentrations was about 29 times greater than would be predicted for flat terrain. The enhanced dilution was found to decrease with increasing downwind distance but increase with increasing stability. Wilson et al. (1976) concluded that under low wind-speed conditions, the ability of the atmosphere to diffuse a gaseous tracer does not decrease toward zero as the atmosphere becomes strongly stable but tends to be bound by some minimum rate of dilution.

Air pollution stagnation, although typically associated with the stagnation of high pressure systems over a region, may at times be due to other meteorological causes. Leahey (1977) indicates that Calgary's (Canada) poor air quality episodes tend to occur in the evening during synoptic situations typified by light south-southwest geostrophic winds and locally mountain-generated northwesterly katabatic winds. The forces that produce the local katabatic winds counterbalance the existing synoptic horizontal pressure gradient forces, leading to a stagnating airmass with a consequential high meteorological urban air pollution potential.

3.2.1.5 Trapping

Trapping is the atmospheric condition that occurs with a persistent upper-level temperature inversion aloft (Hewson, Bierly, and Gill 1961). Figure 3.8 illustrates trapping for different valley situations. In Figure 3.8a, the valley is shallow compared with the height of the stack; lateral diffusion is unimpeded if other conditions are essentially steady. In Figure 3.8b, the valley is deeper but the inversion base is still above the valley sides and

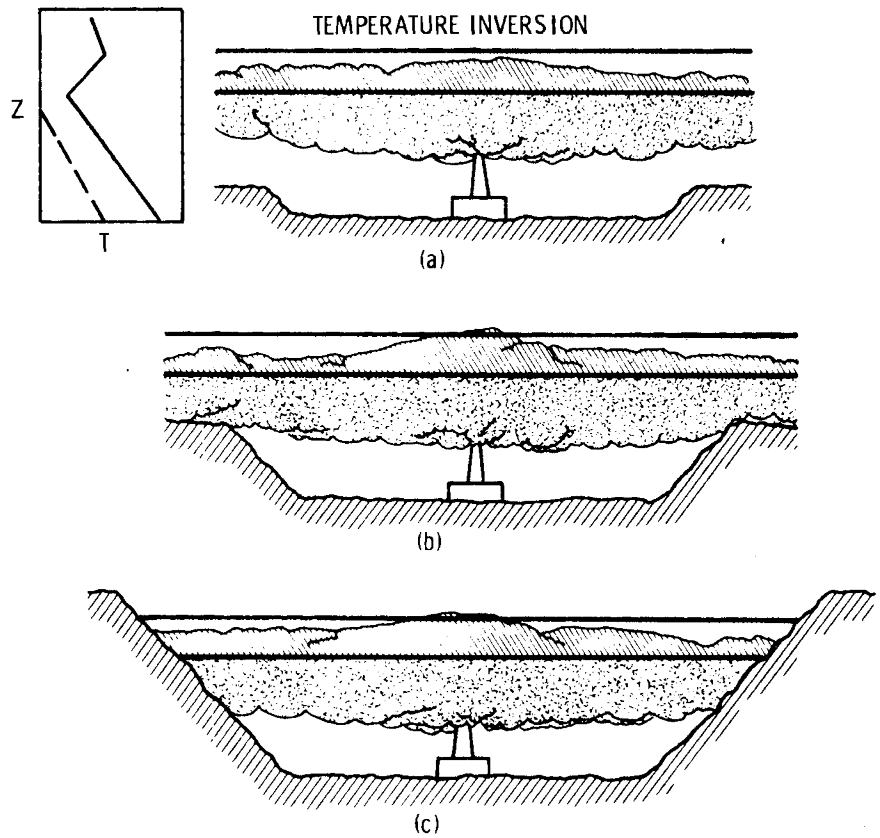


FIGURE 3.8. Trapping of an Effluent Below a Temperature Inversion Aloft and Valley Walls. See text for description (adapted from Hewson et al. 1961)

lateral diffusion is not impeded. In Figure 3.8c, the effluent is confined in the valley by its walls and the inversion aloft. The latter condition could easily produce high effluent concentrations throughout the valley especially with the development of local winds and fumigation conditions.

Smoke or cloud domes are associated with trapping temperature inversions in valleys. Apparently one of the first investigators to report on the phenomenon was Hewson and Gill (1944) in their studies in the Columbia River valley. On clear, calm mornings smoke from stacks would rise above the general elevated smoke layer (100 to 200 m) and then spread out and sink again to become part of the general layer. This formed a hemispherical dome above the smoke layer, 15 to 30 m in thickness. Such domes were also observed by pilots when a uniform cloud layer was over the valley. Scorer (1972) discusses similar phenomena in relation to cooling towers and industrial plants.

Whiteman and McKee (1978 and 1980) have indicated from, at least, a few valleys in central Colorado that deep inversions that build up within the valley during the night descend into the valley during the hours after sunrise. This inversion descent will have important implications for dispersal of any pollutant that is emitted into a very stable nocturnal inversion layer since the material may be dispersed after sunrise during inversion descent.

3.2.1.6 Downwash

Downwash has been mentioned in Section 3.1 and depicted in Figure 3.4. This phenomenon is important for sources located on the leeside of mountain ridges and hills. Scorer (1978); Smith and Kauper (1963); Smith (1965); and Whaley, Lee, and Gaines (1979) have indicated that leeside airstreams such as large amplitude leewaves, rotors, hydraulic jumps, and separation can affect plume behavior by downwash and other terrain effects, such as channeling. Since leewaves, rotors and hydraulic jumps are usually associated with strong winds, plumes are quickly diluted, thus reducing pollutant concentrations in a short distance.

3.2.1.7 Summary

Landforms affect plume transport and dispersion in different ways. Pollutant concentrations at local sites can often be increased with impaction, stagnation, and trapping. Conversely, downwash on the lee of hills or ridges can decrease pollutant concentrations. Channeling of airflow by valleys, ridges, and mountains has a strong effect on plume transport. Currently, models are being developed to simulate many of these phenomena but the results are preliminary since not all of the variables are simulated properly. Many of the physical details are lacking because of limited field data. Also, good field data will be necessary for the eventual verification of modeled phenomena.

3.2.2 Local Winds

Local winds were defined in Section 2.4 as land-sea (lake) breezes, thermal-slope winds and valley-mountain winds. This section reviews the state-of-the-art on how these local winds and terrain affect plume transport and dispersion.

3.2.2.1 Land-Sea (Lake) Breezes

The majority, if not all, of the studies concerning dispersion over or near water bodies pertain to coastal sites or large lakes such as the Great Lakes. The problem of diffusion near or over lakes in a relatively complex terrain site has not, apparently, been addressed. However, a brief examination of past studies should provide valuable insight into the problem of diffusion near large bodies of water, although information on complications resulting from complex landforms will be missing or have to be evaluated in a qualitative manner.

The problem of dispersion and transport near coastlines and large lakes has been considered by Hewson, Bierly, and Gill (1961); Bierly and Hewson (1963); Hewson and Olsson (1967); VanderHoven (1967); Collins (1971); Raynor et al. (1975); Lamb, Lorenzen, and Shair (1978a); Lamb, Shair, and Smith (1978b); Lyons (1975); and Misra (1980). Lyons has written a very detailed review concerning turbulent diffusion and pollutant transport in shoreline environments based principally on observations taken near Lake Michigan. According to Lyons, the problem of dispersion near shorelines can be generally divided into three cases: 1) gradient offshore flow, 2) gradient onshore flow, and 3) the land-sea (lake) breeze.

The first case addresses diffusion over water. When land temperatures are cold compared to those of water, elevated plumes traveling over the water are rapidly mixed or dispersed because of the unstable conditions over the water (Figure 3.9a). When the land temperatures are warmer than those of the water, intense low-level over-water temperature inversions occur during these periods. elevated plumes traveling over the water become stabilized and form a ribbon-like plume that may be transported several kilometers with little lateral or vertical diffusion (Figure 3.9b).

The second case concerns the flow of marine air inland under gradient conditions; that is, without the complicating effects of land or sea breezes. As discussed in Section 2.4.1, the low-level flow reacts to an almost step-function discontinuity in surface roughness, temperature, and evaporation. One of the manifestations of surface changes is the formation of internal boundary

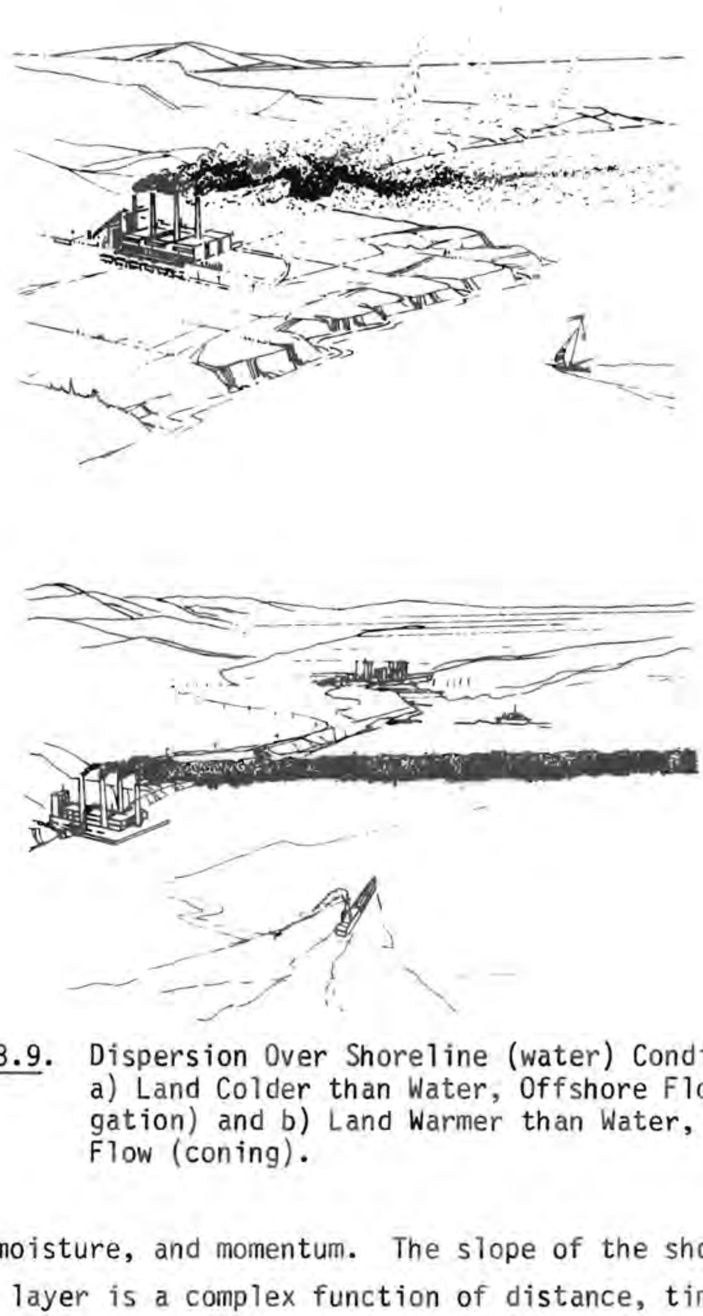


FIGURE 3.9. Dispersion Over Shoreline (water) Conditions: a) Land Colder than Water, Offshore Flow (fumigation) and b) Land Warmer than Water, Offshore Flow (coning).

layers for heat, moisture, and momentum. The slope of the shoreline thermal internal boundary layer is a complex function of distance, time and mesometeorological conditions.

Two shoreline diffusion regions of greatest concern are: a) plume trappings that are more frequent during night or cloudy days, and b) continuous (dynamic) fumigation, prevalent during stable onshore flow on sunny days. Plume trapping is common near shorelines. The thermal stratification resulting in plume trapping may be due to subsidence, warm air advection or modification of a stable layer's lowest portions to a nearly adiabatic state by heating from the surface, and/or mechanical mixing. An example is shown in Figure 3.10a. Shoreline fumigation is the characteristic of elevated sources emitting into stable onshore marine air undergoing intense overland modification as described in Figure 3.10b. Lyons (1975) has summarized the various models for estimating surface concentrations beneath a fumigation zone.[a] The fundamental basis of most of these models is the steady-state Gaussian plume model tending to make the models conservative with respect to predicting pollutant concentrations. The last case considers the effects of land-lake (sea) breezes in transport and diffusion. Lyons and others have shown that shoreline air quality can be seriously affected by lake (and land) breezes. His observations show that there is at least a partial recirculation of gases and aerosols within the lake breeze cell (Figure 3.11a) and suggest that there is a size-sorting phenomenon that allows sub-micron particles to concentrate in great numbers in the shoreline atmosphere whereas the larger particulates fall out over the lake surface (Figure 3.11b). Lyons points out that airflow in the lake breeze is not two-dimensional as most modeling efforts assume. In reality, a distinct long-shore drift may exist and pollutant trajectories resemble flattened helixes. Thus, pollutants released from a point on a shoreline will usually not return to the point of origin.

Fumigation of shoreline tall stack plumes is a frequent occurrence during lake breezes; however, it does not occur during all lake breezes (Figure 3.11c). If the lake breeze has only penetrated inland a few kilometers it may not have sufficient fetch for the thermal internal boundary layer to reach plume level. In this case, the rather undiluted plume will drift inland, rise sharply at the front, and then be advected lakewards aloft in the return flow layer. If a portion of it does not subside into the return-flow layer, when it once again crosses the shoreline, it will be in relatively high diluted concentrations.

(a) For other modeling attempts see Misra (1980) and Meroney, Cermak, and Yang (1975).

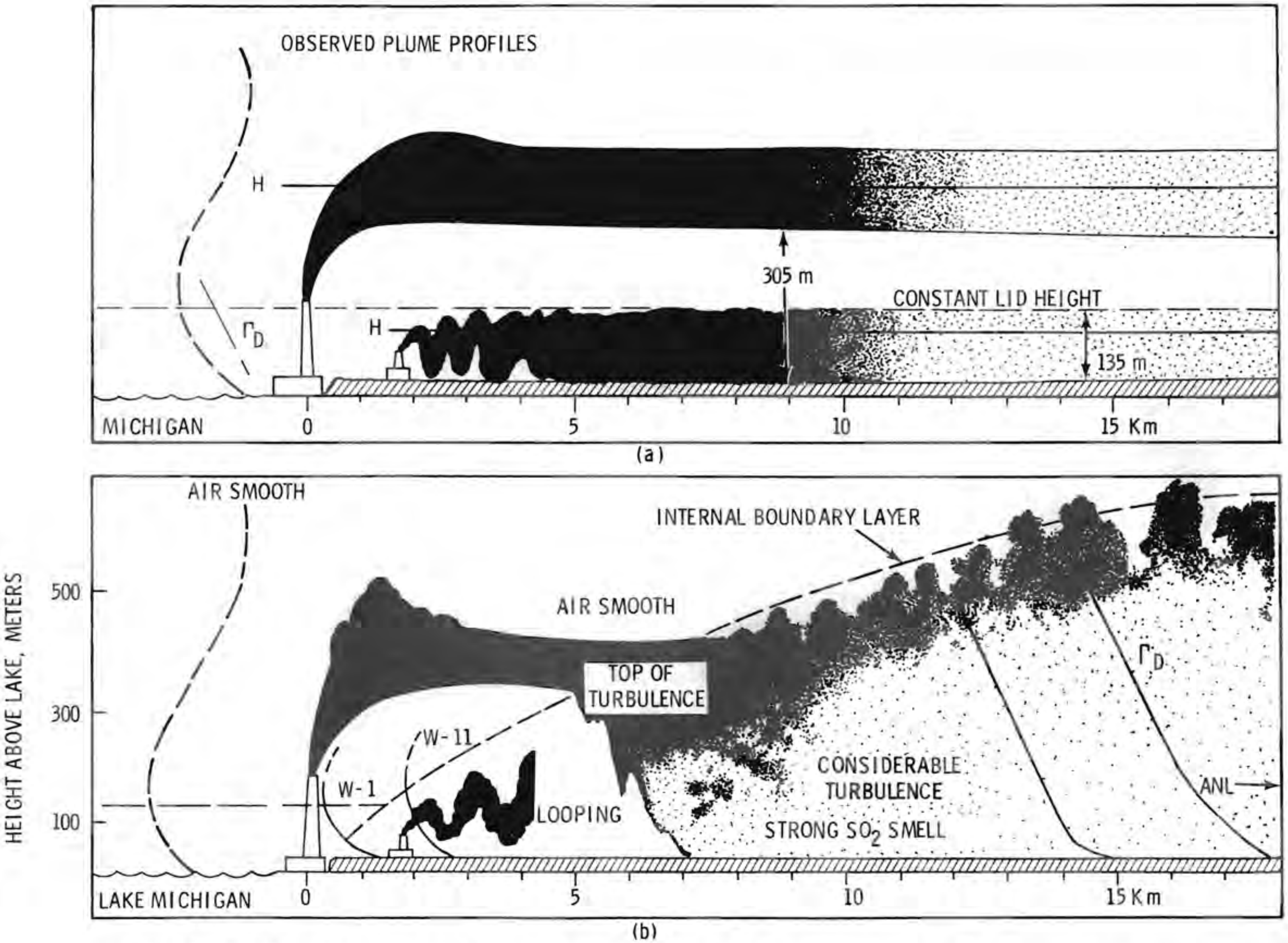


FIGURE 3.10. Dispersion Over Shoreline (land) Conditions: a) Trapping of Plumes Moving Over Land and b) Dynamic Fumigation of Plumes Over Land (adapted from Lyons 1975).

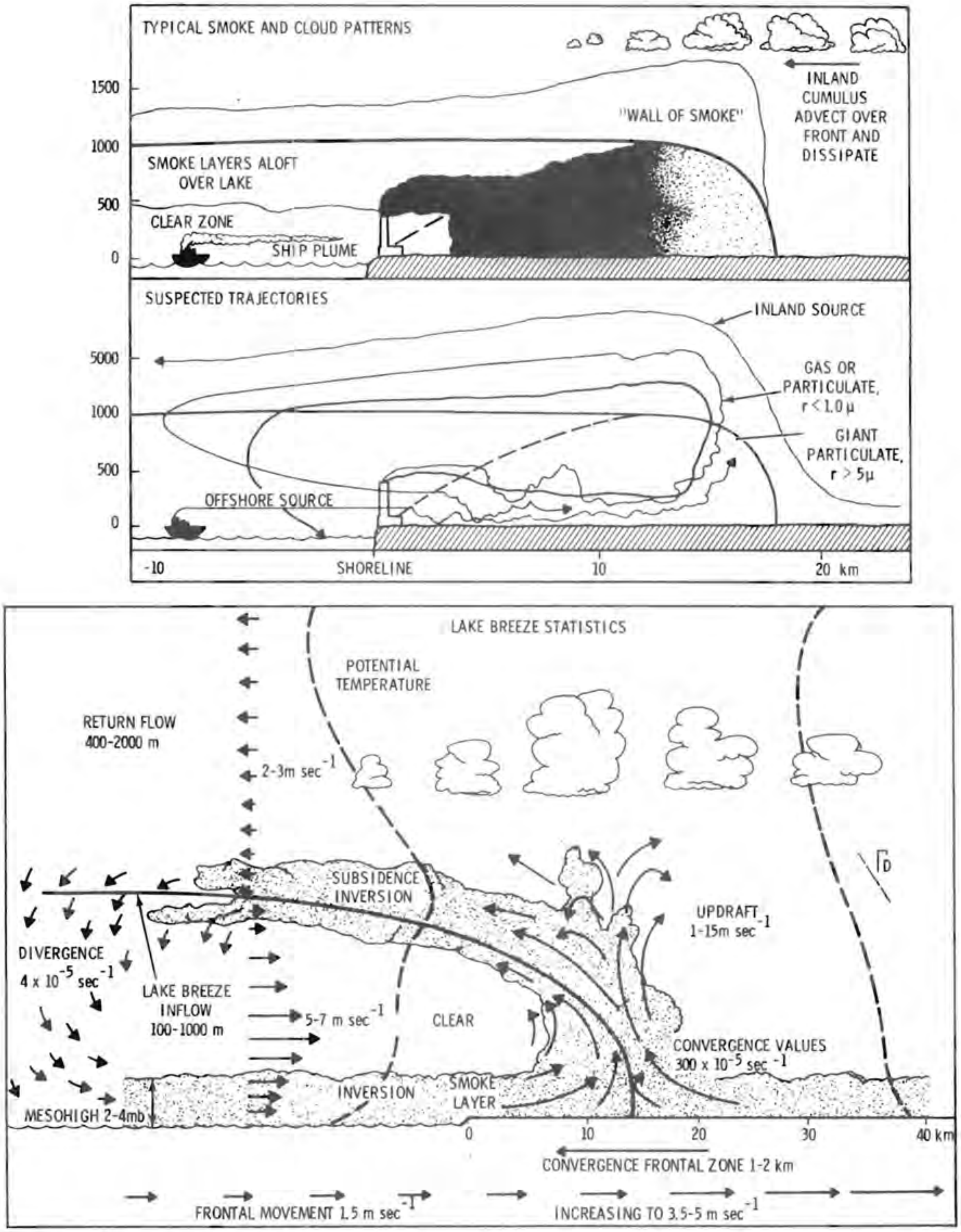


FIGURE 3.11. a) Characteristics of a Typical Lake Breeze During a Period of Light Gradient Winds and b) Streamline and Probable Trajectories of Small and Large Particulates in a Well-Developed Lake Breeze Cell (adapted from Lyons 1975).

The recirculation phenomenon is not a significant factor in terms of producing extremely high surface values of pollutants from individual plumes, although it will add to the regional pollution burden. While surface concentration can be potentially as high as those discussed during fumigation for gradient onshore winds, the effects of strong wind shear on separate plumes are likely to result in somewhat lower total values during lake breezes.

Some of the complications of lake effects associated with complex terrain have been qualitatively evaluated by Hewson and Olsson (1967). In complex landforms, the size, depth, and shape of the lake must be considered. A shallow lake is heated more rapidly than a deep lake. Evidence indicates that a thermal circulation within the lake is developed for lakes with an average diameter of more than 2 km. Patterns around a long, narrow lake in a deep valley or canyon are probably more similar to those of a river then to those of a wide or almost circular lake.

Figure 3.12 illustrates the possible effects on transport and diffusion for a lake adjoined by sloping ground. Figure 3.12a illustrates the daytime conditions when an upslope or valley wind is generated. An upslope wind by day would apparently aid or add to a lake breeze circulation. At night, a downslope wind would aid or strengthen the land breeze (Figure 3.12b).

3.2.2.2 Thermal-Slope and Valley-Mountain Wind Systems

Davidson (1961) evaluated meteorological factors important to transport and dispersion by local valley winds in the valleys of Vermont. These factors included wind direction and speed, stability, and diffusion coefficients. He considered a source located in midvalley under the influence of the nighttime valley-plain (mountain) wind. The relative height of the source with respect to the velocity profile of the valley-plain wind determined the speed at which the pollutant was transported down valley. The height of the valley-plain systems varied from night to night with the average maximum wind speeds on the order of 3 ms^{-1} to 6 ms^{-1}. Thus, the average wind speed at pollutant carrying levels can range from 1 ms^{-1} to 6ms^{-1} depending on the effective height of the source, the height of the ridgeline, and the strength of the prevailing flow.

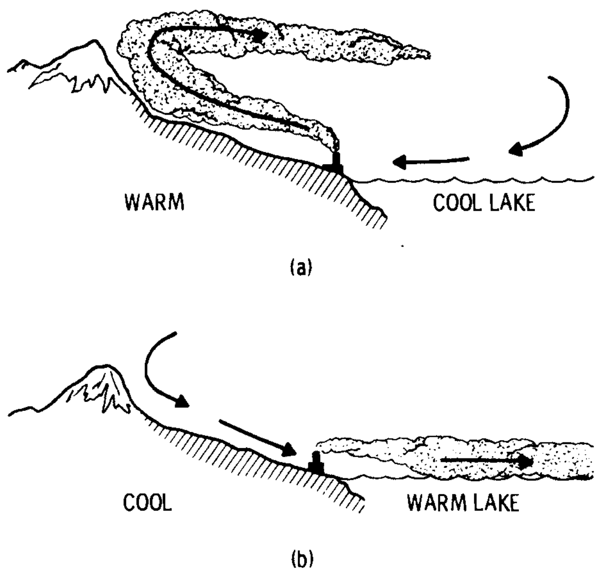


FIGURE 3.12. a) Probable Circulation Pattern and Plume Transport Near a Lake and Sloping Ground during Daytime and b) Same as Above But for Nighttime

Several studies have indicated that except for a relatively thin layer (30 to 100 m) near the slopes of the valley, most of a valley is filled with the valley-plain wind. There appears to be little gradient across a valley but this may depend on the valley. Near the walls of a valley, a shallow layer of air moves downslope on both sides at a speed of about 2 ms^{-1}. The maximum velocity shear occurs near the top of the valley-plain system and near the valley sides. Davidson speculates that if a sizeable pollution source were located on the slopes of a valley and had significant buoyancy and velocity rise, the plume would quickly penetrate the shallow downslope system and join the valley-plain flow.

A qualitative evaluation of transport and dispersion modes in a complex mountain-valley was conducted by DeMarrais, Downing, and Meyers (1968). The evaluation was based on vertical temperature soundings and wind observations in forested terrain in Idaho. Five different modes of transport and diffusion were identified based on moisture, temperature, wind, type of canyon or valley drainage, and time of day. These were:

- Early Morning, Moist Air, V-Shaped Drainage--Transport and dispersion of aerosols under these conditions was difficult to predict. Transport from the area of release was limited because of low wind speeds. Predictions of horizontal dispersion was difficult because of variable winds and vertical dispersion was restricted because of thermal stability. Restricted transport would result in aerosols eventually being deposited at or near the area of release. After sunrise, vertical mixing would bring aerosols down into the trees.

- Early Morning, Dry Air, V-Shaped Drainage--Transport and dispersion generally followed a predictable pattern until about 1 to 1-1/2 hours after daylight first reached the area. Weak drainage winds prevailed. Thermal stability restricted vertical mixing until sunlight reached the area. After drainage wind dissipated prediction of transport and dispersion was difficult. After the sun had reached the treetops, aerosol would mix downward similar to the moist air case

- Afternoon, Dry Air, V-Shaped Drainage--In light winds, prediction of transport and dispersion was difficult. Vertical mixing would bring aerosols to treetop level in a short time. This mixing would extend through the crowns and to the ground. In strong winds, vertical mixing was enhanced by mechanical turbulence and the aerosol would mix through the tree crowns to the ground.

- Evening, Dry Air, V-Shaped Drainage--Transport and dispersion was predictable but elevated aerosols would not reach the ground very fast. Thermal stability kept elevated aerosols from mixing downward for most of the period of darkness.

- Dry Air, U-Shaped Drainage--Transport and dispersion was more predictable during typical mountain-valley winds. Downvalley and downslope winds prevailed until sunlight reached the valley floor; upslope and upvalley winds would persist until the valley floor was shaded Transport was unidirectional most of the day and night.

Elevated released aerosols were brought down into trees in the morning and afternoon but not in the early morning or evening. Vertical mixing was increased after the sunlight reached the valley floor. High concentrations of aerosols would be brought to the surface at this time.

During the past 35 years, a number of diffusion experiments have been conducted in landforms where local valley-slope wind systems were an important aspect in the transport and diffusion processes (See Appendix IV for a list of these experiments or studies). One of the more informative and influential investigations of local winds in valleys and local pollution problems was conducted by Hewson and Gill (1944). Whaley and Lee (1977) recently added to their work.

An important result of Hewson and Gill's study was the definition of two types of fumigation that may occur in valleys: first, those resulting from a temporal (diurnal) change in the wind and turbulence regime associated with the solar heating of a cool ground surface (Type I), and second, those resulting from a spatial change in the wind and turbulence regime associated with the advection of local air masses (Type II).The first type of fumigation (Type I) is common in the early morning hours (08 to 10 AM) and occurs occasionally during the late afternoon (18 to 20 PM). Pollutants from elevated plumes reach the valley bottom nearly simultaneously at all points along the valley. The strength of the fumigation along a valley is roughly proportional to the distance from the source. Late afternoon fumigations are similar but not as well defined.

The second type of fumigation (Type II), according to Bierly and Hewson (1962), results from the low-level heating of air as it passes over an artificial (city) or natural (water) neat source (nondiurnal). Type II fumigation episodes do not have clearly defined characteristics as seen in the Columbia River study. Hewson and Gill indicate that the strength of fumigation was approximately inversely proportional to the distance from the source. The pollutants did not appear simultaneously at the surface all along the valley, but were usually noted first near the source and then in succession at various points down the valley

Tyson (1969) considered a dynamic type of fumigation condition associated with the dissipation of the downvalley mountain wind and onset of the upvalley wind in the Umsinduzi valley near Pietermaritzburg in South Africa. In individual valleys around Pietermaritzburg, the valley wind is often initiated at or near the entrances to the valleys; thereafter, it advances upvalley undercutting the downvalley mountain wind. When the valley wind undercuts the more stable downvalley mountain wind, a marked discontinuity in the wind velocity and turbulence fields produce severe dynamic fumigations as illustrated in Figure 3.13. In contrast to diurnal fumigations, which produce synchronous groundlevel pollution maxima along the valley floor, dynamic fumigation delays the time of maximum groundlevel pollution with increasing distance upvalley.

The interaction of synoptic and local winds can produce some complex dispersion phenomenon. An example was given in Section 2.4.4.1 and Figure 2.28 where a shear zone or micro-front developed in the Snake River basin of Idaho when downslope winds interacted with channeled synoptic scale winds. The effect of this shear zone on an elevated plume from a stack and a surface experimental explosion is illustrated in Figure 3.14. In the case of the elevated plume, the lower part was traveling northward and the upper part southward. However, in the case of the explosion, the lower part of the plume was traveling southward and the upper part northward. Apparently the slope of the shear zone or micro-front sloped in opposite directions for these two different days.

In the 1960s, McMullen and Perkins (1963), Smith and Wolf (1963), Smith and Kauper (1963), and Smith (1965) conducted a series of diffusion measurements over ridges of moderate relief when the prevailing airflow was parallel and perpendicular to the ridgeline. In one study, the winds were nearly parallel to the ridge. Under these conditions, flow over the ridge did not contribute substantially to the results but the influence of the drainage wind system on ground dosage was clearly evident, particularly within the lateral canyons. In three studies, prevailing winds were approximately perpendicular to the ridgeline. Under the influence of organized ridge flow patterns, the highest dosages were observed on the lee slope of the ridge mainly as the

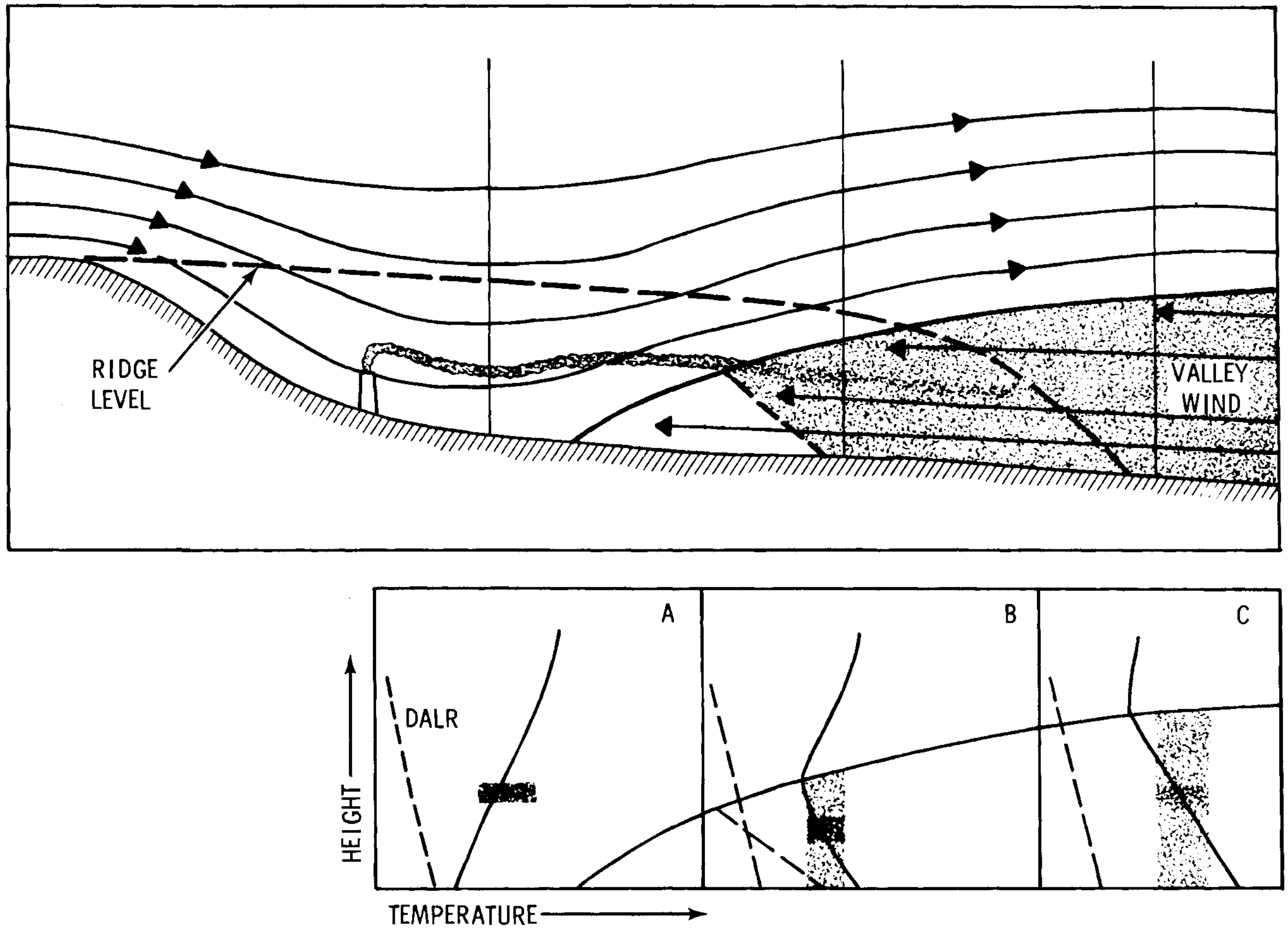


FIGURE 3.13. A Schematic Illustration of Dynamic Fumigation Associated with the Groundlevel Onset of a Valley (daytime) Wind (adapted after Tyson 1968)

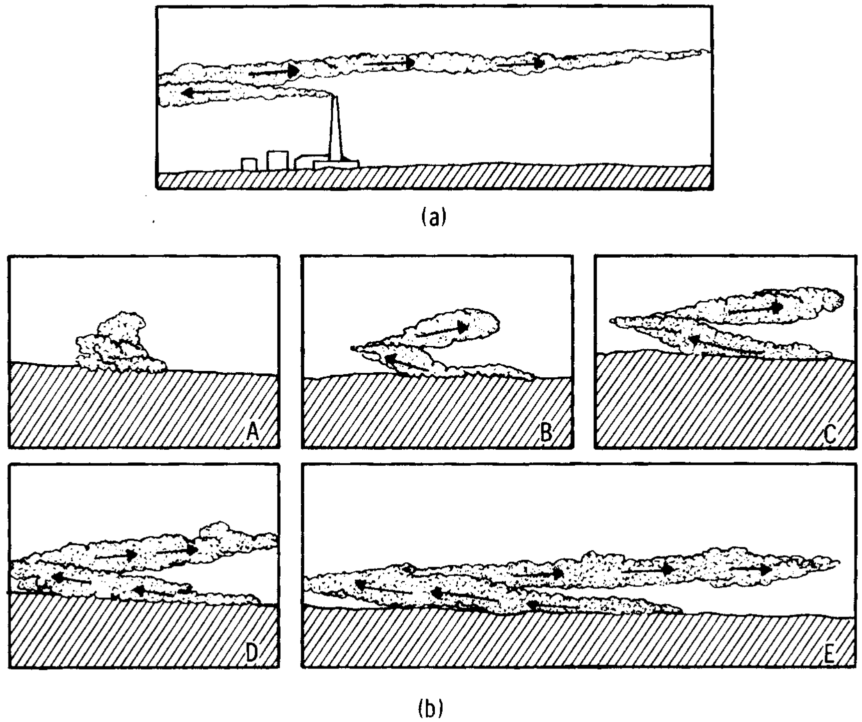


FIGURE 3.14. a) Effect of Shear Zone on an Elevated Smoke Plume from a 75-m Stack, NRTS, Idaho 0722 MST, April 11, 1952 and b) Effect of Shear Zone on a Cloud of Smoke from an Experimental Explosion, August 29, 1945 (adapted from Wilkins 1956)

result of leewave patterns but occasionally with the drainage flow effects superimposed. On the windward slopes, occasionally small dosages were observed due primarily to drainage effects.

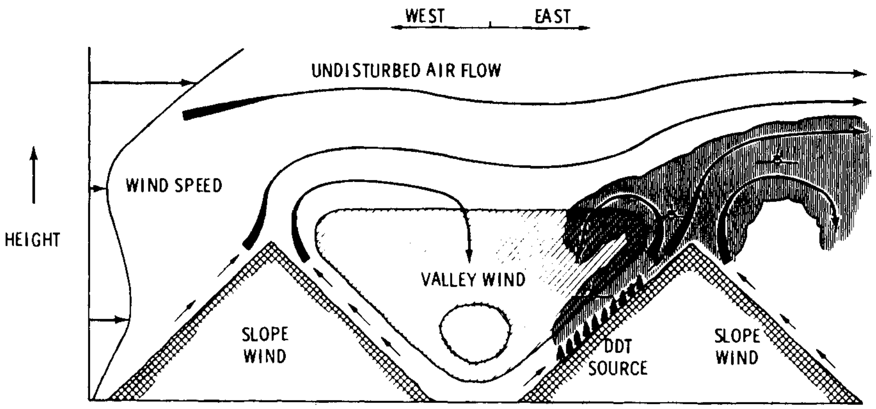
A number of transport and diffusion tracer experiments have been carried out in the 1970s. Details of the experiments are beyond this report but each one emphasizes the importance of local valley-slope wind systems, particularly nighttime drainage winds, in transporting and dispersing polltants within complex landforms.

Classically, the daytime counterpart of the slope-valley wind system consists of a shallow layer of upslope motion adjacent to the valley sides and a wind directed from the plain to the valley in midvalley. Exceptions have already been noted in Sections 2.4.2 and 2.4.3. With the development of thermal instability, the wind regime in a valley may become quite turbulent, in part, because of the effect of the prevailing airflow interacting with the valley system. An effluent plume within the valley may be dispersed fairly rapidly under such conditions.

Various field measurements have shown that daytime valley and upslope winds transport pollutants into mountain areas from distant sources. Edinger et al. (1972) showed that photochemical oxidants and other pollutants formed in the marine layer of the Los Angeles Basin are vented up the slopes and over the crest of the San Bernadino Mountains during the day. Layers of high oxidant concentrations were detected above the inversion base, suggesting that some pollution is vented up the slopes and subsequently advected back to the south. The pollutant transport occurs in an upslope wind layer of about 150-m thickness.

Williams, Brady, and Willison (1977) have provided evidence that numerous ponderosa pine and other tree species in the lower elevations of California are oxidant injured as the result of advection of oxidant pollutants into the forests by upslope and upcanyon flow. Additional studies by Carroll and Baskett (1979) and Unger (1978) have provided more evidence of the transport of photochemical ozone from distant sources into the slopes and valleys of the Sierra Nevadas.

At times, pollutants may be transported out of mountainous areas to other distant regions with the assistance of daytime valley-slope winds. Resuspension and translocation of a pesticide, DDT, from a forested and mountainous area of northcentral Washington was measured by Orgill, Sehmel, and Peterson (1976). Sampling occured on warm days when daytime valley-slope winds were well developed (Figure 3.15). DDT concentration profiles and fluxes indicated that convection, convective turbulence and upvalley and slope winds were predominant mechanism for DDT resuspension and transport. Later, after the spray operation had ceased, DDT concentrations were sampled in rainwater in the eastern states (Peakall 1976).



MODEL OF VALLEY - SLOPE WIND IN AN IDEAL SHAPED VALLEY WITH HYPOTHESIZED DISPERSION OF DDT AND APPROXIMATE LOCATIONS OF SAMPLING AIRPLANE

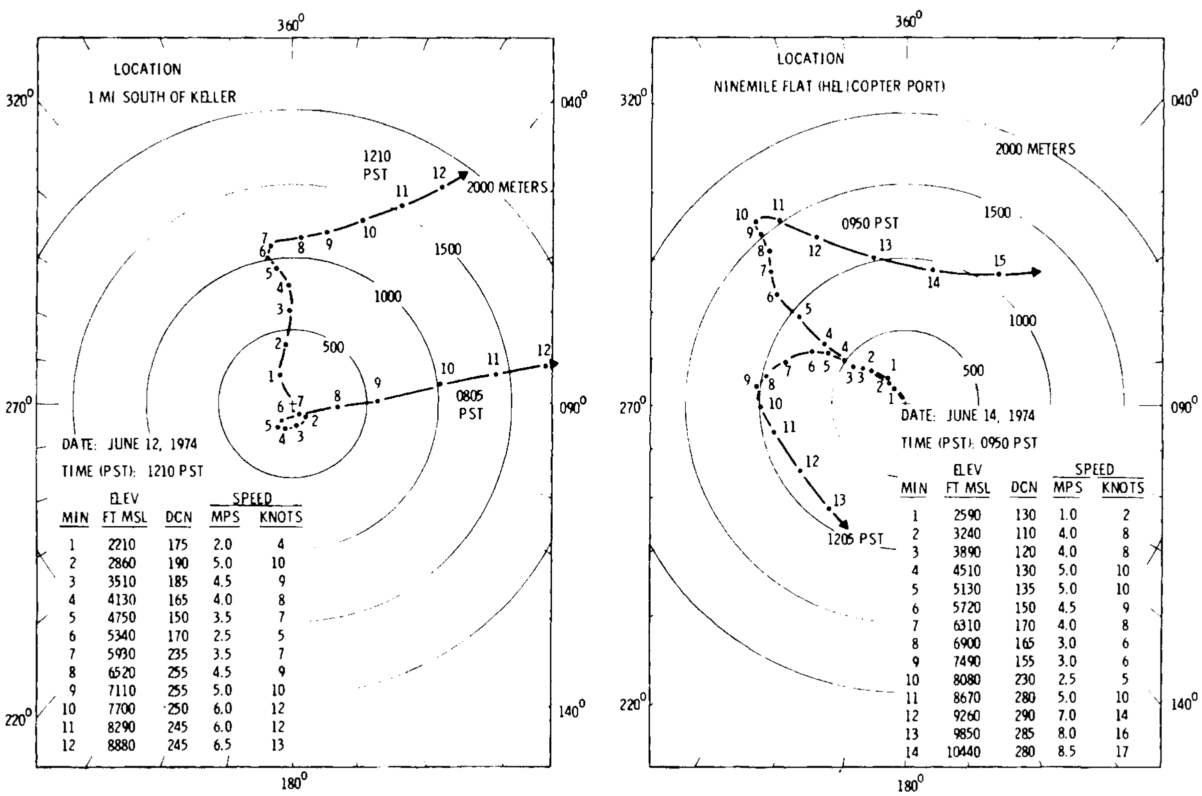


FIGURE 3.15. a) Schematic Illustration of Valley-Slope Wind in an Ideal Shaped Valley with Estimated Resuspension and Dispersion of DDT from the Slopes. Valley wind is from the south and b) Supporting Hodographs from June 12 and 14, 1974 showing the Development of the Valley Wind in the Lower Levels (adapted from Orgill et al. 1976)

3.2.2.3 Summary

Local winds, such as lake breezes and valley-mountain wind systems affect plume transport and dispersion in several ways. Past field studies have revealed some of the complexities that occur when plumes are released in local wind systems. Transition periods of diurnal wind circulation exert a profound effect in the dispersion of plumes as has been revealed by studies in valleys and along lake shores. Interactions between synoptic and local winds often cause marked shear layers which can result in complex plume transport. Many of these effects are still in need of serious study. Modeling of such phenomena is in its infancy but is currently increasing in number. Additional field studies will be needed to supply input and verification data.

3.2.3 Dispersion Coefficients

Four principal approaches are used to theoretically and empirically represent the diffusion or spreading action of a turbulent atmosphere (USAEC 1968):

- Gradient-transport theory or eddy-diffusivity method (Gaussian plume model)
- Statistical methods
- Dimensional analysis or similarity methods
- Practical schemes, such as the Pasquill-Gifford method σ_y, σ_z versus downwind distance method.

Generally none of the above methods are very adequate for predicting diffusion in very rough terrain although a number of modifications in original methods have produced approximate results.

The Gaussian plume model and the Pasquill-Gifford (PG) empirically-determined graphs of $\sigma_y(x)$ and $\sigma_z(x)$ as a function of downwind distance and atmospheric stability have gained wide acceptance as a method for estimating downwind concentrations of atmospheric pollutants (Pasquill 1961 and Gifford 1961, 1968, and 1975). The Gaussian plume model can be expressed in a number of ways depending upon the type of concentration desired (instantaneous, annual average, centerline, crosswind average). For example, the downwind centerline groundlevel normalized concentration from an elevated point source is given by (Turner 1967)

$$\frac{C(x)\bar{u}}{Q} = \frac{1}{\pi\sigma_y(x)\sigma_z(x)} \exp\left[-\frac{1}{2}\left(\frac{H}{\sigma_z(x)}\right)^2\right] \tag{3.2}$$

where

$C(x)$ = air concentration at some downwind distance x

$\bar{u}$ = mean wind speed

Q = release rate

H = effective height of the release.

This equation is applicable until the downwind distance is reached where the top of the plume intersects the top of the mixing layer. At this distance, which can be variable, the equation must be modified or other methods used to calculate the concentration.

Two important parameters in this model are the lateral and vertical dispersion parameters, $\sigma_y(x)$ and $\sigma_z(x)$, respectively. Values generally applied with the Gaussian plume model are obtained from the Pasquill-Gifford graphs of $\sigma_y(x)$ and $\sigma_z(x)$. The PG curves have been applied to a large variety of situations but they were actually intended for use under rather limited circumstances: wind speeds greater than 2 ms^{-1}, nonbouyant plumes, flow over open country, and downwind distance of only a few kilometers (Gifford 1976). A number of studies have attempted to improve the PG curves. Briggs (1973) has proposed a series of interpolation formulas for $\sigma_y(x)$ and $\sigma_z(x)$ that extend the dispersion distances out to 10 to 100 km, but are still applicable to a surface roughness (z_0) of 3 to 30 cm. Smith (1972) has extended Pasquill's scheme for $\sigma_z(x)$ to include downwind distances up to 100 km and a variety of surface roughness. Hosker (1974) used a form of Smith's results to estimate the effect of surface roughness on downwind air concentrations.

Rough terrain is known to increase the dispersion of plumes because of an increase in the mechanical turbulence induced by roughness. In the case of rough terrain, use of the standard PG curves for estimating the impact of a source may lead to an underestimate of the air concentration near the source and an overestimation of concentrations further downwind (Miller 1978, Leahey 1974). Uncertainties in the use of the standard PG curves in rough terrain have promoted a number of field measurement programs for obtaining data on the variation of pollutant concentrations, $\sigma_y(x)$ and $\sigma_z(x)$ with distance for a variety of landforms. Shearer and Minott (1976), Shearer, Minott, and Hilst (1977), Minott and Shearer (1977), Minott, Shearer, and Marker (1977), Draxler (1979) have summarized several of these studies (see Appendix IV for a number of these special diffusion experiments for different landforms).

These studies were conducted under a wide variety of conditions and landforms; but in general, dispersion in complex terrain was found to be greater than that which would be predicted using the standard Gaussian and PG techniques.[a] The effects of complex terrain were most pronounced at night during stable conditions and least during the day under unstable conditions. The enhanced dispersion during stable conditions may be site-specific. For example in the Huntington Canyon study, neutral stability experiments showed five times greater dilution for canyon axial concentrations. Strong inversion tests resulted in canyon plume centerline dilutions fifteen times greater than calculations using the standard PG curves (Figure 3.16). Start, Dickson, and Wendell (1975) suggested that plume dilution was the result of enhanced mechanical turbulence because of gradient winds near the mountain tops, density flows originating in side canyons, and turbulent wakes within the canyon. At Julich, West Germany, which is a rough site but one with homogeneous topography, dispersion during unstable and neutral conditions was enhanced by factors similar to those at Huntington Canyon; but at night, the dispersion at Julich was enhanced only by about a factor of two compared with the factor of fifteen observed at Huntington Canyon.

(a) For another viewpoint see Tank (1976).

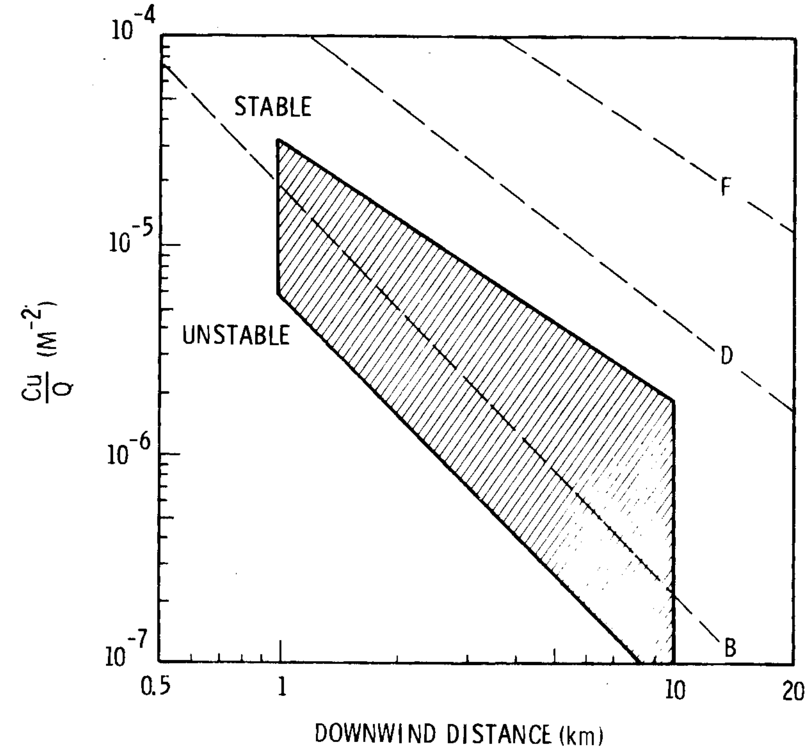


FIGURE 3.16. The Variation of Downwind Normalized Concentration at Garfield and Huntington Canyon (Utah) for Different Stability Conditions. Flat terrain estimates are shown as dashed lines (adapted from Draxler 1979)

The results of some studies (Minott, Shearer, and Marker 1977) indicate that vertical dispersion in deep valleys may differ little from vertical dispersion in less severe terrain. However, other studies indicate that the vertical dispersion can be very dependent on the direction and speed of the prevailing airflow over the ridgeline. Orgill et al. (1971), Wooldridge and Orgill (1978) and Wooldridge (1974), have observed that eddy diffusivity coefficients (vertical, cross-valley, along-valley) measured by constant-volume balloons in two valleys in central Colorado increased in a monotonic fashion up to the levels of the highest surrounding peaks under the influence of moderate gradient flow aloft. Wooldridge (1980) has obtained an expression for the height dependency of the vertical coefficient of diffusivity for momentum in terms of two field meteorological variables: $U_{\star}$ (friction velocity) and $|\vec{V}_g|$ (gradient wind). This expression is

$$K_m(Z) = ZU_\star\kappa \exp \left|-120 \, |\vec{V}_g|^{-0.65} Z^{0.76} U_\star^{-0.11} f^{0.76}\right| \qquad (3.3)$$

where $f = 2\Omega \sin\phi$, κ is the vonKarman's constant = 0.4 and z is height above surface.

A field experiment was undertaken in the Cache Valley of northeastern Utah to compare observed values of the vertical eddy diffusivity coefficient for momentum as a function of height with the values computed from Equation 3.3. Generally, the field results showed reasonable agreement with values calculated through Equation 3.3. Analysis of the behavior of Equation 3.3 shows that values of K_m in the lower part of the profiles respond more to the values of $U_\star$ than to $|\vec{V}_g|$. The upper part of the profile depends more upon $|\vec{V}_g|$, reflecting the effect of momentum fluxes from upper winds on circulations in the planetary boundary layer (Wooldridge 1977).

Horizontal dispersion in valleys depends to a large extent on the width of the valley because valley walls act as a physical barrier to lateral dispersion. Panofsky (1969) and Turner (1967) assume that if the lateral spread of a plume is limited by a valley of width y, then the concentration is assumed uniform across the width of the valley and may be estimated by

$$C = \frac{2Q}{\sqrt{2\pi}\, \sigma_z y u} \exp\left[-1/2 \left(\frac{H}{\sigma_z}\right)^2\right] \qquad (3.4)$$

However, such a result will not work for all weather conditions. Subsequent dispersion studies carried out in deep valley terrain should be directed towards better quantification of the effects of the valley walls on the horizontal dispersion rate.

3.2.4 Dispersion in or Over Forest Canopies

A number of field experiment studies of forest dispersion have been reported but most of these experiments were conducted in forests over rolling or flat terrain (See Draxler 1979). Fritschen et al. (1969) conducted short-period and distance releases inside a forest in the foothills of Mt. Rainier, Washington-Oliver (1973) and Bergen (1975) investigated the airflow within a forest using smoke trails to reveal the short-term horizontal and vertical motion. Wind tunnel studies of tracer diffusion into model forests were reported by Meroney (1968) and Meroney and Yang (1969). Mathematical models of forest diffusion have been developed by Calder (1961) and Ford and Lomen (1969).

Particulate dispersion into and within forests over level terrain have been summarized by Raynor, Hayes, and Odgen (1974) and Raynor et al. (1975). These experiments and other studies have demonstrated that dispersion through a forest differs significantly from that over open terrain. It is suspected that forest diffusion in rough terrain will differ from forest diffusion in level terrain. However, some aspects of forest diffusion in rough terrain may be similar to that in level terrain but without supporting data it will be difficult to identify these features. DeMarrais, Downing, and Meyers (1968) provides some qualitative data with respect to dispersion of elevated aerosols into trees located in valleys during daytime convective conditions (see Section 3.2.2.2).

A forest in complex terrain affects diffusion of tracers depending on: 1) the type and density of the forest; 2) the degree to which the forest alters the temperature, wind, and turbulence; 3) the type of landforms; and 4) the location of release points with respect to the forest. Variations in forest density alter the contribution of mechanical and convective turbulence to diffusion. Variations in crown density alter the thermal properties, and holes in the crown permit sunlight to penetrate to the forest floor. Therefore, seasonal, diurnal, and spatial variations in diffusion are expected.

The location of the source and receptors with respect to the forest is very important. Elevated sources above the forest may be more susceptible to terrain effects than effluent releases in the forest canopy. Releases below the canopy, depending on forest density, stability and wind, may remain for the most part below the canopy. Particulate plumes will lose material by impaction and deposition near forest edges and within the canopy.

Flemming (1967) has considered the effects of forest boundaries and terrain on smoke dispersal particularly in relation to smoke damage to forests on the Erzgebirge ridges. This smoke damage occurred mainly at levels over 650 m above sea level and apparently the smoke comes from sources 20 to 30 km away. Flemming discusses the general meteorological conditions, local winds, and turbulence that result in smoke plumes impacting on the ridge and forest.

3.2.5 Moderate-to Long-Range Transport and Diffusion

Diffusion experiments reviewed in the previous sections were studies in which measurements were primarily made in the range of 5 to 50 km. Sampling beyond a few tens of kilometers becomes very expensive because of the complicated logistics, increased cost, and the large amount of tracer that has to be released to overcome the background concentrations. These experiments are difficult in level terrain and become much more difficult and complicated for complex terrain. These difficulties face both the experimentalist in the operation and interpretation of his field trials, and the theoretician in attempting to formulate a realistic mathematical scheme of prediction.

In the moderate-to-long range (20 to 100 km) diffusion problem, the properties of the entire planetary or Ekman boundary layer are involved, with the changes in wind direction, speed, and turbulence intensity with height being important factors. Roughness or terrain may also change the diffusion process. Topographically-induced eddies or turbulence caused by landforms will exhibit a wide range of scale sizes. Those eddies of scales comparable to, or less than, the dimension of an actual effluent plume will contribute to plume diffusion; the effect of the larger size eddies will simply be the advection of the plume *in toto* along a meandering course (Figure 3.17).

In general, for long-range dispersion, wind shear is more important than turbulence. Pasquill (1962) qualitatively assessed the role of along-wind and cross-wind vertical wind shear in enhancing dispersion observed at a particular level. He summarized the process in two steps. First, shear produces spread when vertical mixing is slow; second, the shear-induced spread is later

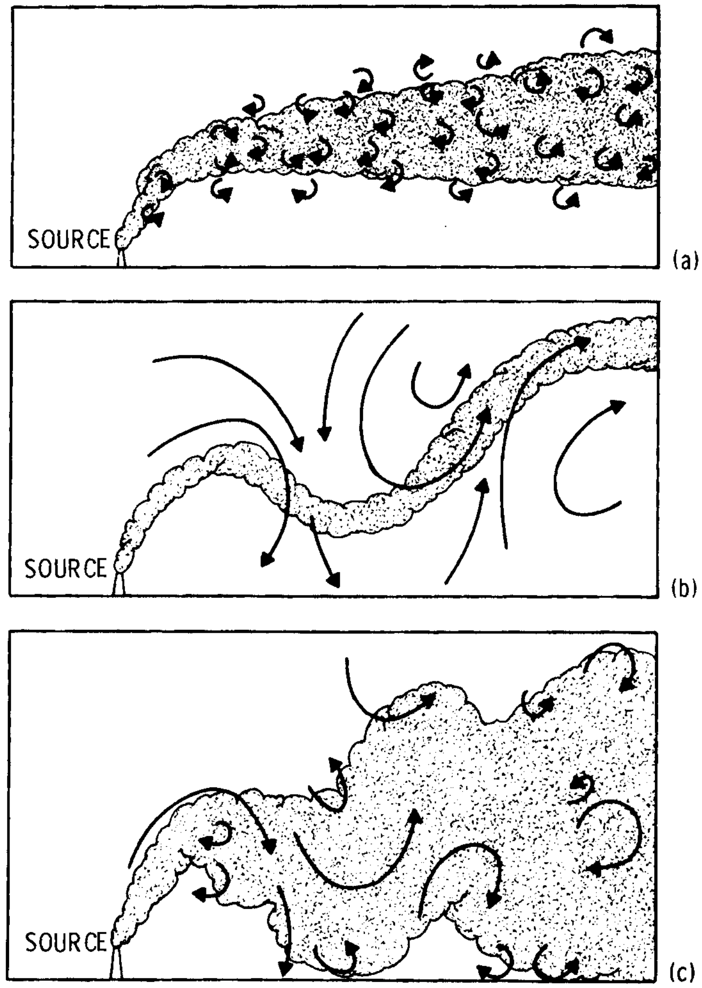


FIGURE 3.17. Idealized Dispersion Patterns: a) a Large Plume in a Field of Uniform Small Eddies, b) a Plume in a Field of Uniform Large Eddies, and c) a Plume in a Field of Different Sized Eddies

observed at a given level when mixing becomes vigorous. Pasquill (1969) examined shorter range diffusion data and determined that the shear-induced dispersion becomes important at about 12 km from ground sources and at about 5 km from elevated sources.

Stability plays an important role in determing the effects of wind shear. Pasquill (Kumar, 1979) indicates that at distances as far as 130 km downwind over rural terrain, very little wind effect is observed during the day when good mixing occurs in the boundary layer; however, during inversion conditions, significant shearing effects occur beyond 2 to 3 km downwind from a source. Numerical results from Kumar (1979) indicate that cross-wind shear effects on horizontal dispersion cannot be ignored beyond 5 to 10 km for a typical stable atmosphere.

A number of investigators such as Leonard (1959), Hilst (1961), Saffman (1962), Pasquill and Smith (1971), Tyldesley and Wallington (1965), Smith (1965), Gee (1967) Pasquill (1971), and Csanady (1972) have shown theoretically that the effect of wind shear on dispersion is considerable at times. Numerical results of Tyldesley and Wallington divide diffusion into four regimes as shown in Figure 3.18: 1) a simple turbulent diffusion region, 2) a plume growing into a shear layer, 3) a transition zone, and 4) a final zone where an upper boundary limits the shear effect. Csanady (1968) has considered, in a theoretical sense. the relative effects of homogeneous roughness and wind shear in the changing of plume dimensions over irregular terrain. Results suggested that for a given roughness the rate of cloud growth at large distances may, because of the wind shear effect, be considerably greater than would follow from an extrapolaion of diffusion data obtained in the surface layer. Wallington and Tyldesley (Smith et al. 1966) indicate that failure to include the wind shear effect as part of our normally acceptable concept of transport by mean motion may well drive dispersion hypotheses into unnecessary complications, but the neglect of the (wind) shear can lead to inadequate interpretation of experimental data.

A number of medium- to long-range diffusion experiments have been reviewed by Draxler (1979) but only three of these experiments may have involved complex terrain. Heimbach, Super and McPartland (1975) presented dispersion data for an elevated release of silver iodide tracer over the terrain near Colstrip, Montana. The extreme limits of these data, "A" and "E" stability, are shown in Figure 3.19. Multiple tracer releases were made at the Idaho National Engineering Laboratory (INEL) and were summarized by Ferber and List (1973) and Clements (1979). The results of SF_6 sampling data are shown in Figure 3.19. Very long-range data stratified by stability are not shown in

REGION	APPLICABLE DISTANCE	PLUME WIDTH
UPPER BOUNDARY INEFFECTIVE	<0.63D	$0.48 \frac{\Delta V'}{u} x$
TRANSITION ZONE	> 0.63D <2.66D	$0.48 \frac{\Delta V}{u} x$
UPPER BOUNDARY LIMITS SHEAR EFFECT	> 2.66D	$4.3 (2K_e x/u)^{1/2}$ $K_e = \frac{(\Delta V)^2 L^2}{120 K_z}$

*FROM TYLDESLEY AND WALLINGTON (1965)
$\Delta V'$ - WIND SHEAR OVER DEPTH 0.58h
ΔV - WIND SHEAR OVER DEPTH h

$$D = \frac{\bar{u} h^2}{2K_z}$$

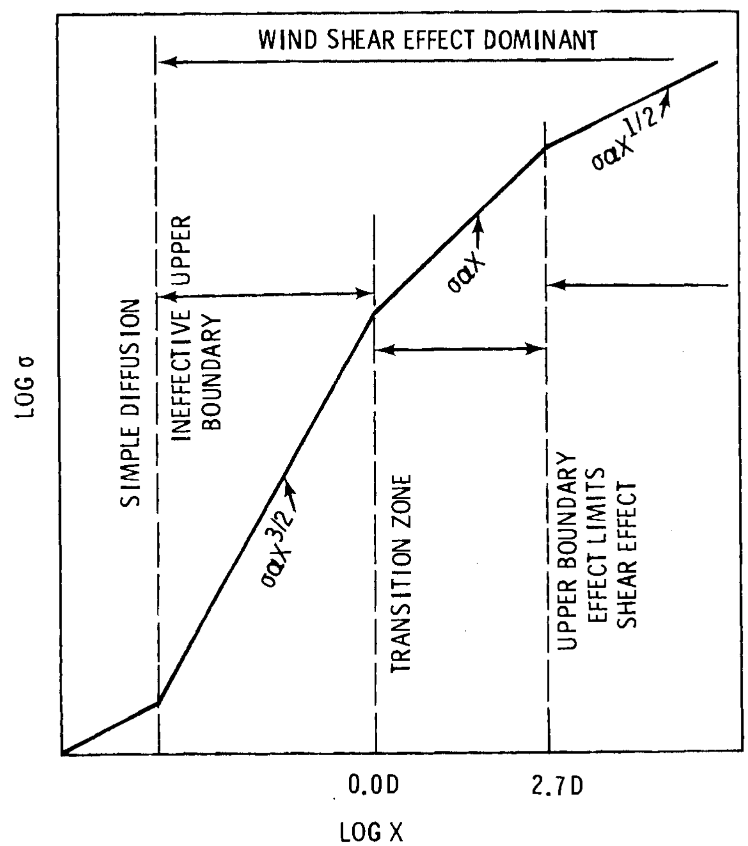


FIGURE 3.16. Schematic Illustration and Table of the Four Phases of Shear-Influenced dispersion (adapted from Tyldesley and Wallington 1965)

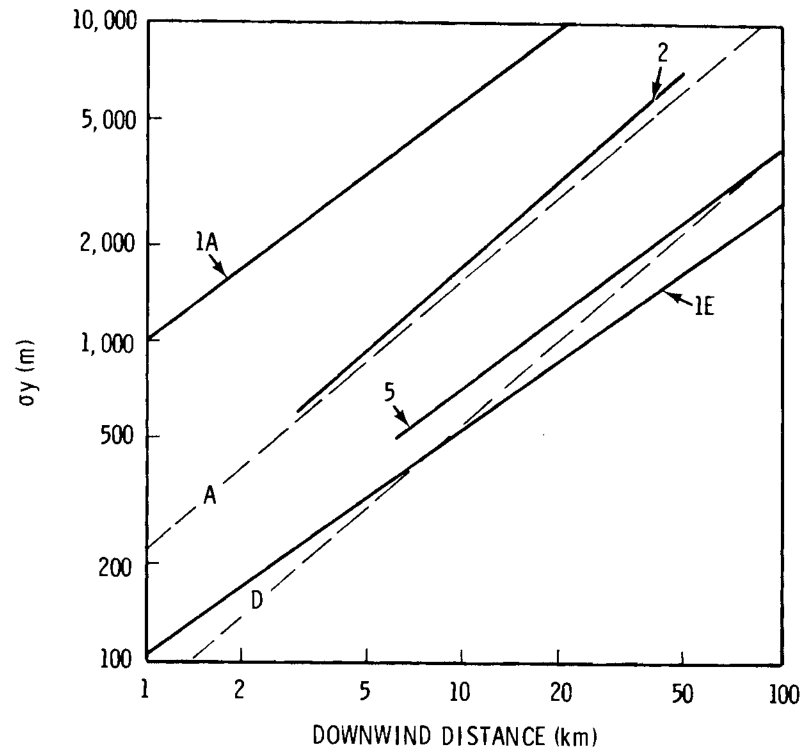


FIGURE 3.19. Long-Range Observed Horizontal Spread (σ_y) for Diffusion Experiments Over Relatively Complex Terrain: 1A-Heimbach et al. (1975) Stability "A" Curve; 1E--Stability "E" Curve; 2--Clements (1979); 5--Ferber and List (1973); Dashed Lines--Pasquill-Gifford Short-Range "A" and "D" Stability Curve (adapted from Draxler 1979).

Figure 3.19 beyond 100 km downwind because the significance of a single stability is hard to evaluate when travel time approaches one-half the diurnal cycle. Other studies (Crozier and Seely 1955; Kassander 1959; Langer, Rosinski, and Edwards 1967; Auer, Veal, and Mariwitz 1968; Willis 1968; Orgill, Cermak, and Grant 1971; Super 1974; and Reid 1979) involved with the transport and dispersion of silver iodide in complex terrain have provided some quantitative data on concentrations and plume dimensions for distances ranging from 35 to 60 km during daytime stability conditions.

The possibility of using three-dimensional tetroon trajectories and dispersion statistics as approximations of three-dimensional air parcel trajectories and dispersion has been under investigation for several years (Angell 1961, Pack 1962; Angell 1964; Angell 1962; and Angell and Pack 1961, and 1965). The tetroon, a superpressured constant-volume balloon, tends to float along a surface of constant density, but is easily displaced from that surface by vertical air motions. Tetroon trajectories closely approximate mesoscale circulation features but their accuracy in amplitude and phase differences between balloon and air motion depend on atmospheric stability, balloon characteristics, the period of motion, and vertical velocities (Booker and Cooper 1965, and Hanna and Hoecker 1971).

Low-level tetroon experimental flights have been useful for indicating air trajectories, terrain influences and dispersion statistics over some 80 km (Dickson and Angell 1968, Angell et al. 1966 and 1972, and Leahey and Hicklin 1973). Tetroon experiments in rugged terrain present a variety of tracking problems especially over long distances. A few experimental flights in the rugged terrain of central Colorado demonstrated strong leewave motions to the lee of ridges at times and separation eddies at other times. Relative dispersion, as indicated by squared total three-dimensional separation rates for dual balloons as a function of time, was proportional to t^3 and t^4 even at the later stages of the flights. Negative separation rates also occurred and were apparently due to cross-valley helical rotors in the valley (Orgill, Cermak, and Grant 1971) (Figure 3.20). However, observation of flights were limited to daylight hours.

Certain orographic-dynamic effects such as deformation of airflow, channeling, impaction of plumes and thermal effects, such as fumigation, and local winds, can play an important role in the long-range transport diffusion problem. A number of different scenarios involving these various effects can be envisioned. An example is shown in Figure 3.21. An effluent plume (or pollutants) emitted in one valley during the nighttime eventually drifts to a second valley where daytime convection fumigates the plume within this valley.

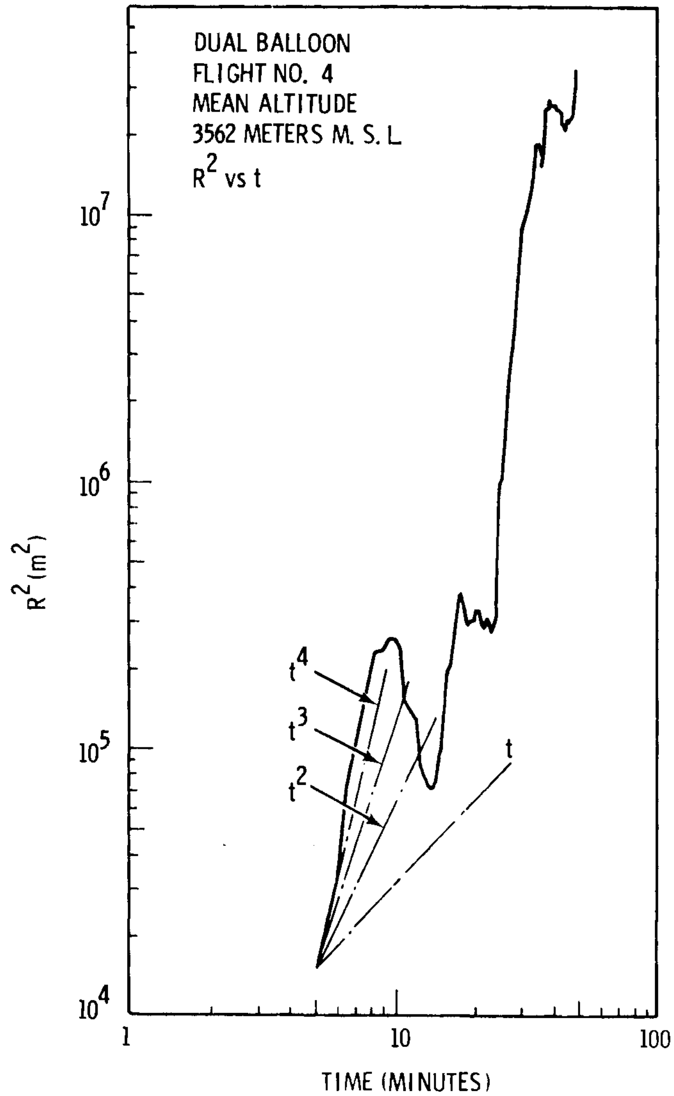


FIGURE 3.20. Squared Separation Rate of Dual Balloon Flight No. 4, Release from Camp Hale, Colorado, as a Function of Flight Time. Rates for t, t^2, t^3, and t^4 are shown for comparison (from Orgill et al. 1971)

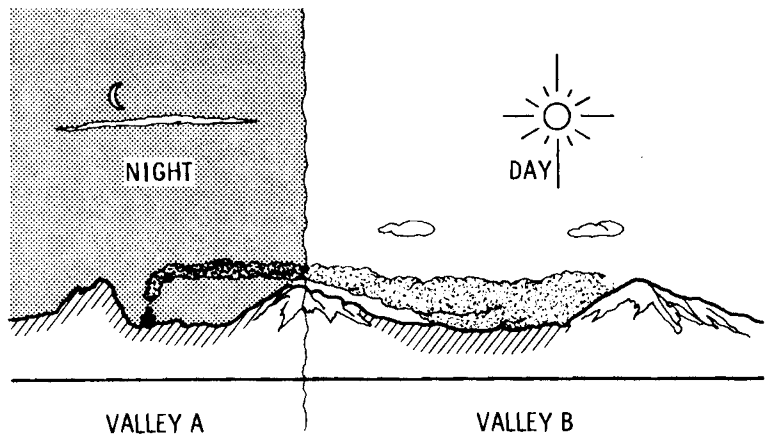


FIGURE 3.21. Schematic Illustration of a Possible Long-Range Transport Problem: A Plume Released in Valley A During the Nighttime is Transported by Winds to Valley B. During the daytime, the Plume fumigates Valley B.

3.3 DEPLETION AND RESUSPENSION FACTORS

This section summarizes the processes responsible for transforming and removing pollutants from the atmosphere and the role complex terrain may play in modifying or augumenting these processes. Complex terrain is envisaged as possibly being both passive and active in this process. In a passive role, certain landforms may alter wind, temperature, and humidity conditions that may augment or decrease transformation and removal mechanisms. In an active role, terrain and, particularly its associated vegetation, may actually participate in the process by acting as a source or sink for pollutants.

Vegetation in the form of alpine forests or even desert shrubs and trees can emit gaseous chemicals to the atmosphere, including hydrogen, a number of low-molecular weight hydrocarbons, several aldehydes, a wide variety of essential oil components, and acidic and basic compounds. Robinson and Robbins (1968) indicated that the major sources of H_2S, NH_3, N_2O and hydrocarbons are natural emissions possibly from plant foliage. Rasmussen (1972) has shown that several forest species release appreciable quantities of volatile organic substances to the atmosphere. The major compounds emitted are monoterpenes (C_{10}) like α-pinene, β-pinene, and limonene and the hemiterpene (C_5) isoprene. Rasmussen and Went (1965) have shown that in temperate forest and field atmospheres the concentration of organic substances such as isoprene and pinenes in the air showed a diurnal variation and were related to the mass of the viable foliage.

Rasmussen (1972) conducted an inventory of North American forest regions for the frequency of occurrences of isoprene and α-pinene released from different tree species and found that 15 was the lowest value for a specific forest-type that emitted terpenes to the atmosphere. More commonly, 100% of the trees of a given forest-type emitted terpene to the atmosphere. An average of 70% is typical of a United States forested regions. Western type forest groups that represent 26% of the total U.S. forest area were primarily α-pinene emitters (~98%) except for the hardwood species (aspen), which were ~100% isoprene emitters. There is a 6.2-fold greater emission level of these hydrocarbons than from manmade sources. The fate of these gaseous olefins in the atmosphere remains largely undetermined.

3.3.1 Chemical and Physical Transformations

Anthropogenic polluting sources vary widely in type but most involve some type of a combustion process in which large amounts of innocuous gases such as carbon dioxide, nitrogen and oxygen and smaller amounts of other gases are released into the atmosphere. The most common gases are: sulfur dioxide, aldehydes and other hydrocarbons. In addition, particulates and trace elements are in the effluent. A more complete classificaton of typical pollutants is shown in Table 3.2 (Altman and Dittmer 1966).

When various pollutants are released in the atmosphere they then become subject to chemical and physical transformations. Chemical transformations are generally of two basic reaction types: thermal and photochemical. In the thermal-reaction category are included a spontaneous breakdown of single reactants and reactions that occur solely because of collisions between two (or more) interacting species. In the photochemical category are included all reactions that occur (either directly or indirectly) through absorption of light in the visible and ultraviolet ranges. Substantial overlap exists between these categories. Physical transformation of atmospheric aerosols involve coagulation, condensation (or its reverse) and nucleation (Hales 1975).

TABLE 3.2. Classification of Air Pollutants

Major Classes	Subclasses	Typical Members
Inorganic gases	Oxides of nitrogen (NO_X)	Nitrogen dioxide, nitric oxide
	Oxides of sulfur (SO_X)	Sulfur dioxide, sulfuric acid
	Other inorganics	Ammonia, carbon monoxide, chlorine, hydrogen fluoride, hydrogen sulfide, ozone
Organic gases	Hydrocarbons	Benzene, butadiene, butene, ethylene, isooctane, methane
	Aldehydes, ketones	Acetone, formaldehyde
	Other organics	Acids, alcohols, chlorinated hydrocarbons, peroxyacyl nitrates, polynuclear aromatics
Aerosols	Solid particulate matter	Dusts, smoke
	Liquid particulates	Fumes, oil mists, polymeric reaction-products

The most common air pollution problems caused by chemical and physical transformation of pollutants are photochemical oxidants and transformation of SO_2 to sulfate These are especially important in long-range transport. Under the influence of sunlight, nitrogen oxides combine with gaseous hydrocarbons to form a complex variety of secondary pollutants called photochemical oxidants. Those oxidants, together with solid and liquid particles in the air, make up what is commonly known as smog. The photochemical oxidant family of pollutants includes, among others, ozone, an unstable, toxic form of oxygen; nitrogen dioxide; peroxyacyl nitrates; aldehydes; and acrolein. In the air, they can cause eye and lung irritation, damage to vegetation, offensive odor, and thick haze. Went (1960, 1965) has proposed that natural terpene emissions can also undergo a photchemical polymerization process that results in aerosols (Aitken nuclei) that are believed responsible for the natural blue haze associated with vegetation, particularly with forests in mountainous terrain.

Topographic landforms favorable for the production of photochemical smog are open and closed basins and long valleys. These topographic features have large depressions within which accumulated pollutants can stagnate and when combined with plentiful sunlight, low-wind speed (poor ventilation), elevated inversion layers and high temperatures create a huge chemical reaction cell in which the photo and other chemical processes may occur. An obvious example is the Los Angeles basin.

Another common air pollution transformation problem occurs with sulfur dioxide (SO_2). Most of the sulfur-burning fuels appear in the resulting exhaust gases as sulfur dioxide and sulfur trioxide. Despite the comparatively small amount of sulfur trioxide produced, its process is important because it drastically raises the dew point of the pollutant gas and readily forms sulfuric acid mist with atmospheric moisture. Under some atmospheric conditions this mist converts a usually colorless plume to a conspicuous bluish-white haze. After many hours in the atmosphere, all of the SO_2 emitted is first oxidized to SO_3, then H_2SO_4 which, in turn, is eventually converted to sulfate. Ammonium sulfate aerosols are suspected of decreasing the scattering extinction coefficient or the visual range in the lower atmosphere (Garland 1969, Eggleton 1969, and Stevens et al. 1980).

Particulates, nitrogen dioxide, photochemical oxidants, and sulfates all affect the visual quality of the lower atmosphere (Charlson and Ahlquist 1969, Samson and Ragland 1977, and White 1976). Particulates, however, are the major reason for visibility impairment. Particles (ash, carbon, dust, and liquid particles) discharged directly to the air scatter and absorb light, reducing contrast between objects and their backgrounds. Particles are also formed by photochemical reactions and by the conversion of sulfur dioxide to sulfuric acid mist. Whenever sulfur pollution is significant (whenever large amounts of coal and oil are burned), visibility diminishes as relative humidity rises (Georgievsky and Rozenberg 1973, Kasten 1969).

During the past few years, interest has increased concerning the impact of air pollution on the quality of scenic vistas, particularly in areas of the western United States where spectacular scenery is enhanced by excellent visibility. The Clean Air Act stipulates the restoration and protection of visibility in natural parks, wilderness areas, and forests (so-called mandatory Class I areas). However, there is some concern over what effects future energy developments in the western states will have on visual range since there is already some indication that visibility is and will be deteriorating in some western states (Trijonis 1978; Pueschel et al. 1978; Leonard, Wecksung, and Williams 1978; and Latimer and Bergstrom 1979).

3.3.2 Dry Deposition

The mechanisms of dry-deposition include: a) gravitational settling, b) turbulent and Brownian diffusion, c) inertial-impaction effects, d) diffusion-phoresis effects, and e) electrical effects. In addition to the above removal effects, deposited material may be re-released to the atmosphere by desorption (in the case of gases) or by mechanical resuspension (in the case of particulates). The above mechanisms are affected by the type of pollutant material, the type of deposition substrate, and by meteorological conditions (Hales 1975).

Chamberlain (1953) surmised that the net effect of these various processes could best be summarized by assuming that the deposition flux (F_d) is proportional to the air concentration measured at some convenient height Z_d:

$$F_d = -V_d \, C(Z_d). \tag{3.4}$$

The resulting proportionality constant (V_d) has the dimensions of a velocity and is called the deposition velocity. Deposition velocities for a variety of substances and atmospheric conditions are available (VanderHoven 1968). Sehmel and Hodgson (1979) have developed a predictive model for correlating particle and gas removal rates from the atmosphere by dry deposition. A review of the state-of-the-art can be obtained from ERDA Symposium Series 38 (Engelmann and Sehmel 1976). However, there appears to be little data on dry deposition processes and deposition velocities in complex terrain.[a]

(a) For a recent contribution see Elias and Davidson (1980).

Studies in the Black Hills of South Dakota (Davis et al. 1976, and Haggard and Davis 1977) indicate that vegetation filters out particulates from the atmosphere. Results show that the Black Hills area is a significant source of clean air and that observed low levels of particle concentrations are primarily the result of precipitation and in-cloud scavenging, elevation variation, and the particle removal mechanism of vegetation (green-area effect). The green area effect alone may serve to reduce ambient Aitken particle concentrations by as much as 50%.

Vegetation is also a sink for certain gases. Hill (1971) and others have suggested from laboratory studies that vegetation could be an important sink for HF, SO_2, CL_2, NO_2, O_3 and to a lesser extent PAN during the growing season. Field observations indicate that HCL is removed rapidly by vegetation because the Cl^- content of vegetation is much higher near an HCL source. Although NO is removed slowly by vegetation, it is converted to NO_2, CH_3NO_3, etc. in the atmosphere, and it would be taken up rapidly in these forms. Although the cleansing action of vegetation may be an important factor in the cycling of certain pollutants, it is not always desirable. Accumulation of fluoride, SO_2, O_3, NO_2, PAN, Cl_2, etc may cause plant injury, although the pollutants are converted to less toxic compounds rather rapidly (Edinger et al. 1972; Williams, Brady, and Willison 1977; Unger 1978; and Carroll and Baskett 1979).

The role of dry deposition in the long-range transport of SO_2 and sulfate is very important for complex terrain. Model studies indicate that SO_2 in the absence of rain and chemical reactions may have a mean travel distance on the order of 700 km when vertical diffusion is limited by an inversion at a height of about 1 km. Background concentrations traveling over such distances are reduced in magnitude by a factor of about four. The effective deposition rate for long-distance travel is largely governed by the diffusive resistance of low turbulence regions adjacent to the surface. This rate is fairly insensitive to details of surface roughness. Overall deposition velocities for SO_2 are around 1 cms^{-1} or less. In the absence of rain the concentration decays exponentially with a decay or travel distance equal to the inversion height multiplied by the ratio of mean wind speed to deposition velocity. When conversion to sulfate aerosol takes place, the corresponding lifetimes and decay distances are increased by an order of magnitude (Scriven and Fisher 1975a and b).

3.3.3 Precipitation Scavenging

In-cloud scavenging of particulates and gases should become an important topic of research in areas of complex or mountainous terrain especially if energy development expands in the western states. The understanding of precipitation scavenging processes has been hampered by the complexity of the phenomena involved and also by the relatively large cost of related measurement programs, which has limited their number and extent.

The delivery of pollutant to the surface by in-cloud scavenging occurs by four individual processes:

- transport of pollutant into the cloud system
- mixing of pollutant within the cloud system
- capture of pollutant by the cloud hydrometeors (rainout and snowout)
- delivery of precipitation-bound pollutant to groundlevel (washout).

The work in the scavenging field has been summarized by Engelmann and Slinn (1970), and Hales (1972 and 1975). An important aspect of precipitation scavenging within clouds is the problem of acid rain (Likens et al. 1979, and Dochinger and Seliga 1975). Measurements of the acidity of rain and snow reveal that in parts of the eastern U.S. and in Europe, precipitation has changed from a nearly neutral solution 200 years ago to a dilute solution of sulfuric and nitric acids today. The main reason for this trend is the rise in emissions of sulfur and nitrogen oxides to the atmosphere with the increase in burning of fossil fuels.

Acid precipitation has existed for many decades in the vicinity of large cities and industrial plants such as smelters, but the phenomenon is now much more widespread. In large areas of the eastern U.S., southeastern Canada and western Europe, the annual average pH of precipitation ranges from 4 to

4.5[a]. Figure 3.22 shows the general area of the eastern U.S. that has been affected by acid precipitation. Evidently similar areas are found farther north in Canada and possibly in the Rocky Mountains (Likens et al. 1979 and Lewis and Grant 1980). The majority of acidic rain and snow falls on New England, New York, and Pennsylvania.

The gradient of acidity is accentuated in mountainous areas that are generally downwind of urban and industrial sources. Mountains enhance precipitation, which continuously removes pollutants from passing masses of air. This is the situation in southernmost Norway and is the case in the Appalachian and mountains of New England in this country. A clear picture of the extent of acid rain is lacking because no consistent monitoring network has been in existence in this country. Acid rain can harm freshwater lakes, streams and growing vegetation.

3.3.4 Summary

A number of significant studies have addressed the problems of chemical transformation and removal, dry deposition, resuspension and precipitation scavanging. In some cases, the effects of complex landforms and their vegetation have been included in these studies. The transformation and removal processes are very important in long-range transport problems including visibility and acid precipitation. As a whole, the information on transformation and removal processes within complex landforms is rather limited. Additional data is needed on dry and wet removal processs for different landform types. Also, an evaluation of the relative importance of these processees in complex terrain is needed.

At the present time, transformation and removal processes are being parameterized into sophisticated computer models (for an example see Latimer and Bergstrom 1979). In some cases, thse models use complex landforms as lower boundary conditions. Yet, until a better data base is acquired on

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(a) A common measure of acidity is pH, which is defined as the negative logarithm of the hydrogen-ion concentration. The pH scale ranges from 0 to 14, with a value of 7 representing a solution that is neutral, values below 7 indicating greater acidity and values above 7 indicating greater alkalinity. Values lower than 5.6 are considered as acid precipitation.

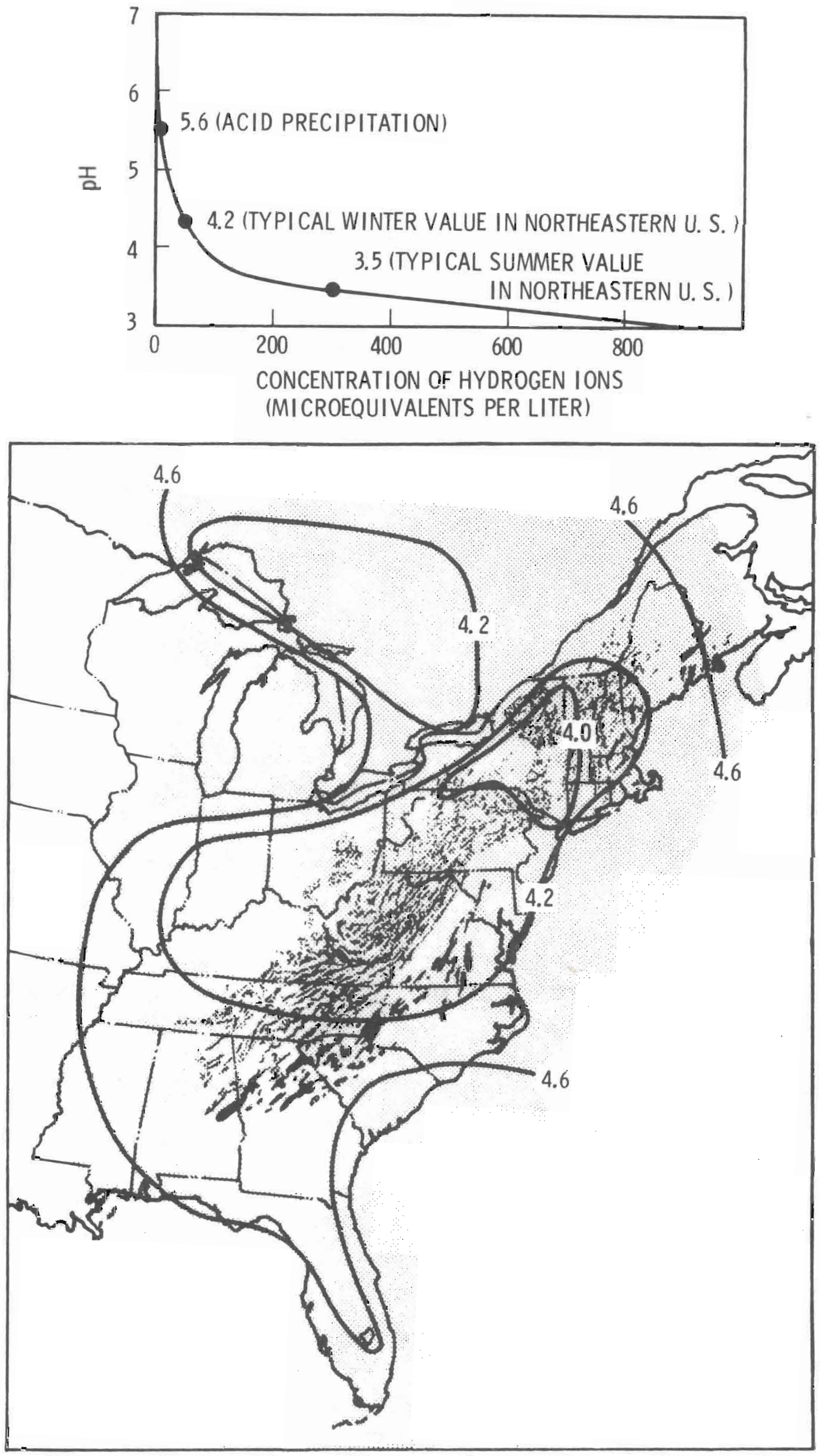

FIGURE 3.22. Major Area in the United States Affected by Acid Rain. Isolines represent 1975-76 averages of the pH (adapted from Likens et al. 1979)

transformation and removal proceses there will probably be an insufficient basis for evaluating the accuracy of numerical models. In the future, modeling and field measurements will have to combine in a joint effort in order to accurately assess the importance and effects of transformation and removal processes in complex landforms.

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4.0 OTHER ASPECTS OF TERRAIN INDUCED AIRFLOW PHENOMENA

4.1 LANDFORMS AND TYPES OF AIRFLOW AND DIFFUSION PHENOMENA

In this section some of the other important aspects of complex terrain problems are briefly examined such as, the relationship of terrain-induced airflow and diffusion phenomena to specific landforms, time and space scales, and the technical difficulty of observing and modeling such phenomena.

Figure 4.1 attempts to associate particular generic landforms with particular airflow or diffusion phenomena that were discussed in Sections 2.0 and 3.0. The dark squares indicate that a particular airflow and diffusion phenomenon is primarily associated with that particular landform. Lighter squares (diagonal lines) indicate that an association between the landform and airflow phenomenon probably exists but more observations are needed to clarify this relationship. An empty space indicates that there is probably poor association or none at all between the landform and the airflow phenomenon.

The landform types and airflow and diffusion phenomena are segregated into various groups such as dynamic and kinematic effects, local phenomena, boundary layer effects, eddies, and plume effects. Turbulence is not included since it is common to all types of landforms in one form or the other. The figure simply attempts to match certain generic landform types to specific airflow and diffusion phenomena. There is not any special meaning in the pattern of the figure.

4.2 TIME AND SPACE SCALES

Landforms such as hills, ridges, valleys, and mountains exist on different horizontal and vertical scales. Consequently, their effects on atmospheric motions and associated weather also exist on different horizontal and vertical scales, and for different time scales. Figure 4.2 classifies terrain effects, or modifications of atmospheric motions, according to an atmospheric scale devised by Orlanski (1975) and illustrates that terrain effects occur over a wide spectrum of length and times scales. The list in Figure 4.2 is used primarily for illustrative purposes and is by no means comprehensive; nor is it meant to imply that terrain effects with similar time or horizontal space scales have dynamic or kinematic similarity.

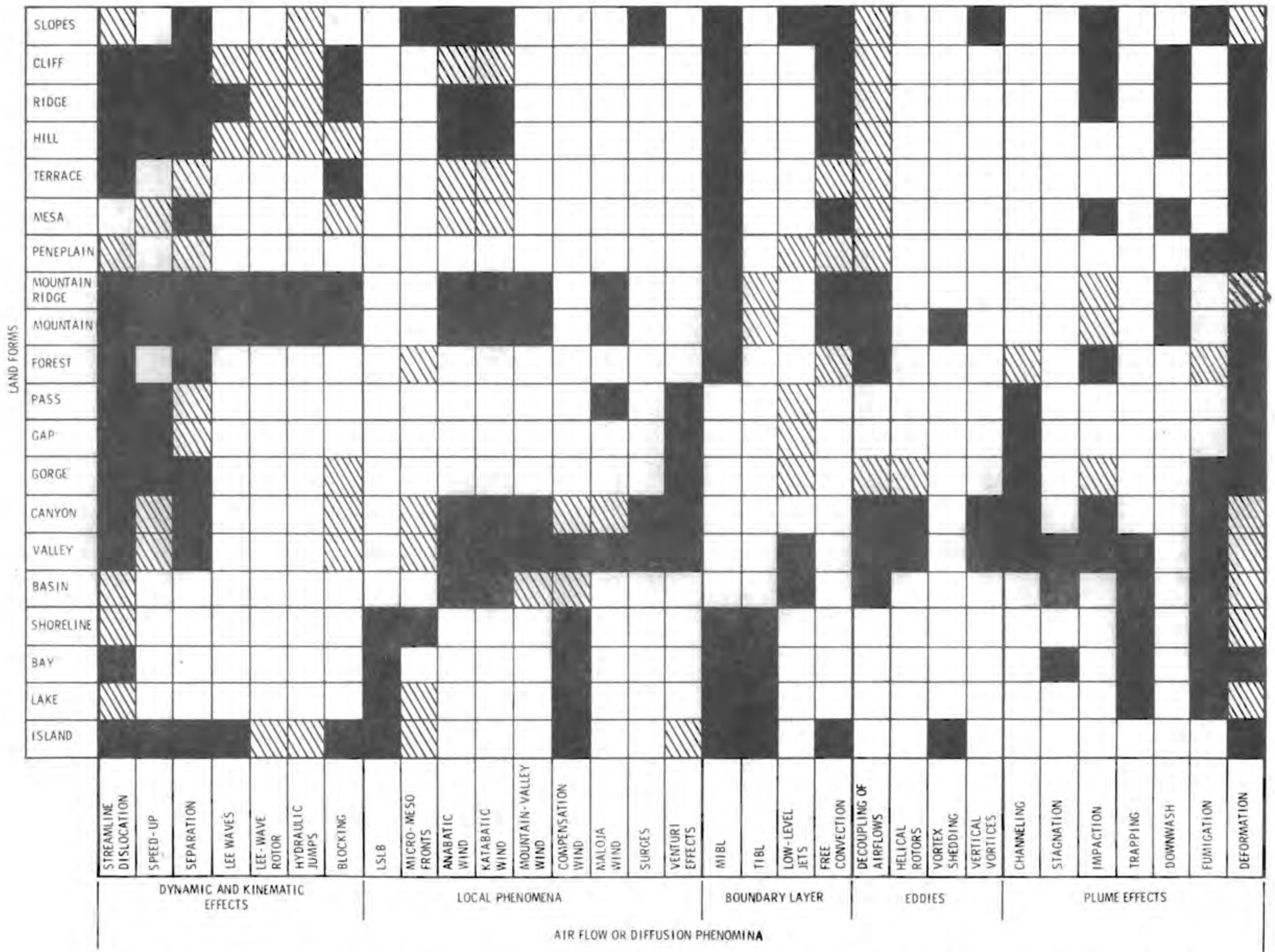

FIGURE 4.1. The General Association of Particular Airflow and Diffusion Phenomena to Different Landforms (see text for explanation of figure)

m H_s	Km^2 A_s			T_s: MONTH — DAY — HOUR — MINUTE — SECOND
	10^8	MACROSCALE α	GLOBAL MOUNTAIN AREA	SUBTROPICAL, JET STREAMS GLOBAL WIND PATTERNS LONG-WAVE RIDGES AND TROUGHS
	10^6	MACROSCALE β	CONTINENTAL MOUNTAIN AREA	MONSOONS; STORM TRACKS CYCLONES AND ANTICYCLONES
3000	10^4	(SYNOPTIC) MESOSCALE α	REGIONAL MOUNTAIN AREA	AIR MASSES FRONTS CYCLOGENESIS
	10^2	MESOSCALE β	MOUNTAIN-VALLEY (PLAIN) BASIN ISLAND	THERMO-TIDAL WINDS LEE WAVES SLOPE-VALLEY WIND VALLEY-PLAIN WIND CHANNELING; EDDIES
300	1	MESOSCALE γ	HILLS RIDGES GORGE CANYON	BLOCKING AIRFLOW SPEED-UP WAKE EFFECTS CHANNELING CANYON, WIND; EDDIES
200	10^{-2}	MICROSCALE α	CLIFFS MESAS TERRACES GAP	AIRFLOW SEPARATIONS AND WAKES CHANNELING; EDDIES
30	10^{-4}	MICROSCALE β	CLIFFS LARGE ROUGHNESS ROUGHNESS TREES	AIRFLOW SEPARATION VERTICAL WIND PROFILES TURBULENCE; EDDIES
		MICROSCALE γ	TREES VEGETATION SMALL ROUGHNESS	VERTICAL WIND PROFILES TURBULENCE; EDDIES
		ORLANSKI CLASSIFICATION	GENERAL LANDFORM OR ROUGHNESS	CLIMATOLOGICAL SCALE; SYNOPTIC AND PLANETARY SCALE; MESO SCALE; MICROSCALE

FIGURE 4.2. A Classification of the Effects of Terrain on Atmospheric Motions (A_S = Area, T_S = Time Scale, H_S = Height Scale)

As Figure 4.2 indicates, various landforms can influence atmospheric motions on all scales, from the scale of the long-wave ridge and trough through lee and wake phenomena to small turbulent eddies. As indicated from the discussion in Sections 2.0 and 3.0, terrain features affect atmospheric motion in four broad categories: 1) through boundary-layer effects as the result of the surface friction (roughness) and local pressure gradients; 2) by the barrier or orographic-dynamic effect caused by blocking of airflow, spatial pressure differences, and the excitation of internal gravity waves and related phenomena; 3) through thermal effects (heating and cooling) caused by vertical and spatial temperature gradients; and 4) through channeling, deforming, stagnating of airflow patterns as a result of valleys, basins, canyons, ridges, and gorges. At any particular site in complex terrain all four of these effects will operate either together or separately; the importance of a given effect will change from day to night, day to day and perhaps month to month, depending on the larger scale airflow patterns and terrain complexity.

4.3 OBSERVING AND MODELING

In this section an attempt is made to rank the technical difficulty of observing and modeling various terrain-induced airflow and diffusion phenomena. Technical difficulty, in a general sense, is defined as the logistical and instrumental difficulty of observing a particular airflow or diffusion phenomenon in field observations, or the difficulty of modeling or constructing a model that will simulate the physics of the phenomenon. Obviously this definition of difficulty is qualitative and is subject to interpretation.

Four methods have been used to study and explain terrain-induced airflow and diffusion phenomena. These are: 1) theoretical or analytical, 2) field experimens, 3) physical modeling (wind tunnel, stratified water channel) and 4) numerical simulation. As Figure 4.3 shows, these four methods are ranked according to relative technical difficulty (scale 0 to 10) for a number of specific terrain-induced airflow and diffusion phenomena. The grouping of these phenomena is similar to Figure 4.1 with the exception of turbulence phenomena.

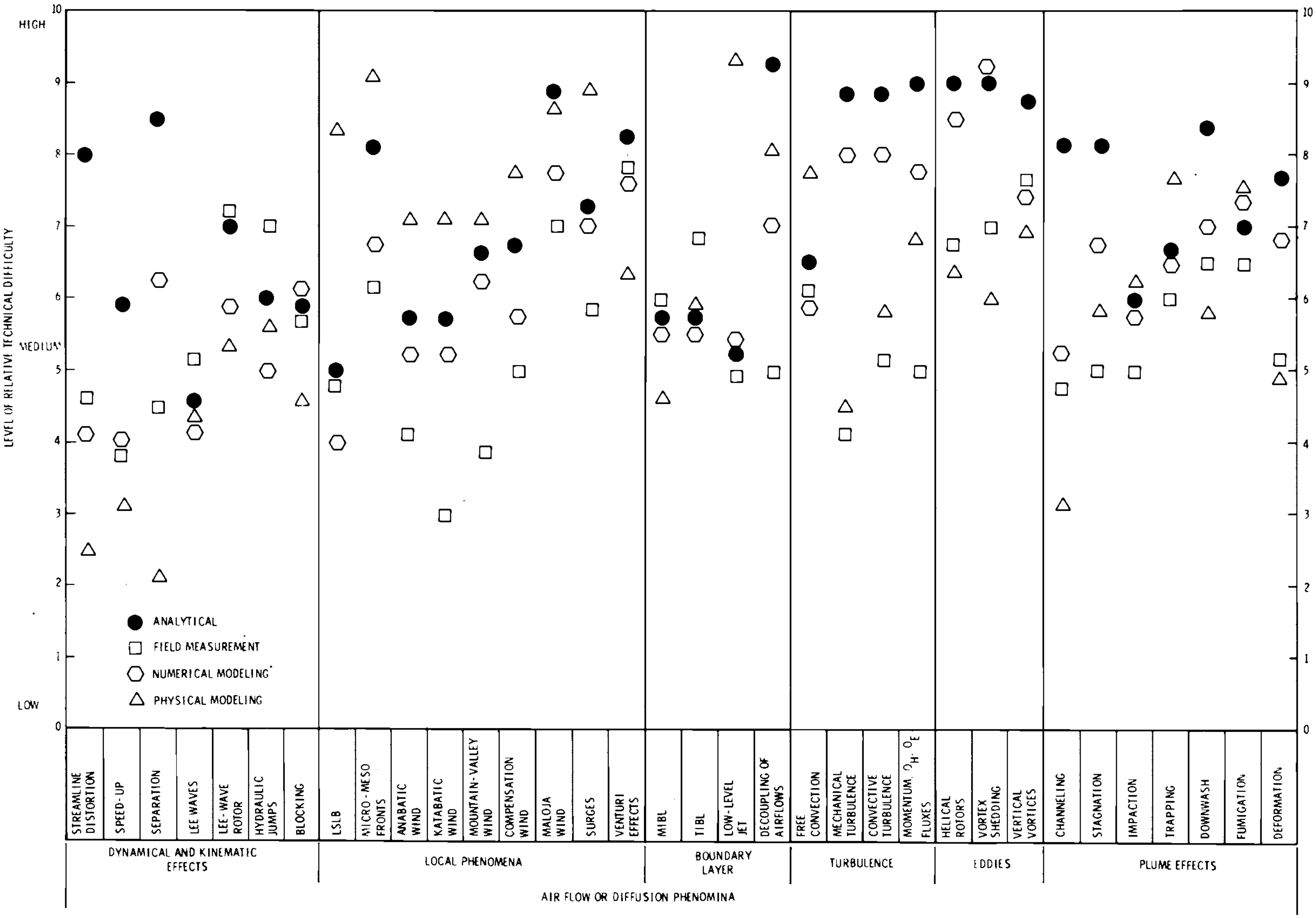

FIGURE 4.3. Qualitative Estimate of Relative Technical Difficulty of Observing and Modeling Various Terrain-Induced Airflow and Diffusion Phenomena

The actual ranking of the various methods with the respect to a specific airflow phenomenon was done by a consensus collected from a small group of the author's colleagues that have had some exposure to complex terrain problems. The points in the figure are essentially the average of the rankings as determined by this mini-survey. For some categories, rankings or opinions of this small group were very close but in some categories they were not. Whether a larger sample of opinions would make the survey any more meaningful may be questionable.

The results from Figure 4.3 are very qualitative but may still provide some information regarding the difficulty of observing and modeling specific terrain-induced airflow and diffusion phenomena. This information may be of some value in defining future problems to be studied in the long-range planning aspects of ASCOT.

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5.0 PRELIMINARY SURVEY OF FUTURE CANDIDATE SITES

Alternative energy sources that satisfy present as well as future environmental standards are desirable. Among the resources strongly considered as alternative energy sources are oil shale, geothermal, and coal gasification/liquefaction. Much of this energy development will occur in rough terrain, either in the West or East; therefore a number of candidate sites for ASCOT may be available. In this section a preliminary examination is made of some possible future sites concentrating on areas where either oil shale, geothermal or coal gasification/liquefaction developments are existing or planned. Sites associated with other energy development activities are acknowledged but not surveyed in depth.

5.1 OIL SHALE SITES

Oil shale is one of the most abundant, but under-developed energy sources in the United States. Oil shale is not shale, and it does not contain oil. It is a fine-grained, compact, laminated sedimentary rock that contains kerogen, an organic high-molecular-weight mineraloid of indefinite composition. Kerogen can be extracted from oil shale by retotring, which results in a hydrocarbon liquid similar to natural crude oil that in turn can be processed and refined much as petroleum.

The major deposits in the United States are found in the Colorado-Wyoming-Utah area known as the Green River formation, and the Upper Mississippian deposits in the East. A map of the Green River formation is shown in Figure 5.1. To process the rich Green River oil shale deposits, different retorting techniques have been developed such as ex-situ (surface), in-situ (below surface), and modified-in-situ. The eastern deposits are of low yield and gasification is considered the most probably technology for recovery of this energy resource.

There are approximately 13 sites in the Green River oil shale region where there has recently been or is present activity to extract the shale (Table 5.1 and Figures 5.2 and 5.3) (Parker 1979 and EPA 1979). However, many potential sites are still in an undeveloped state. Site-specific information is

GREEN RIVER FORMATION

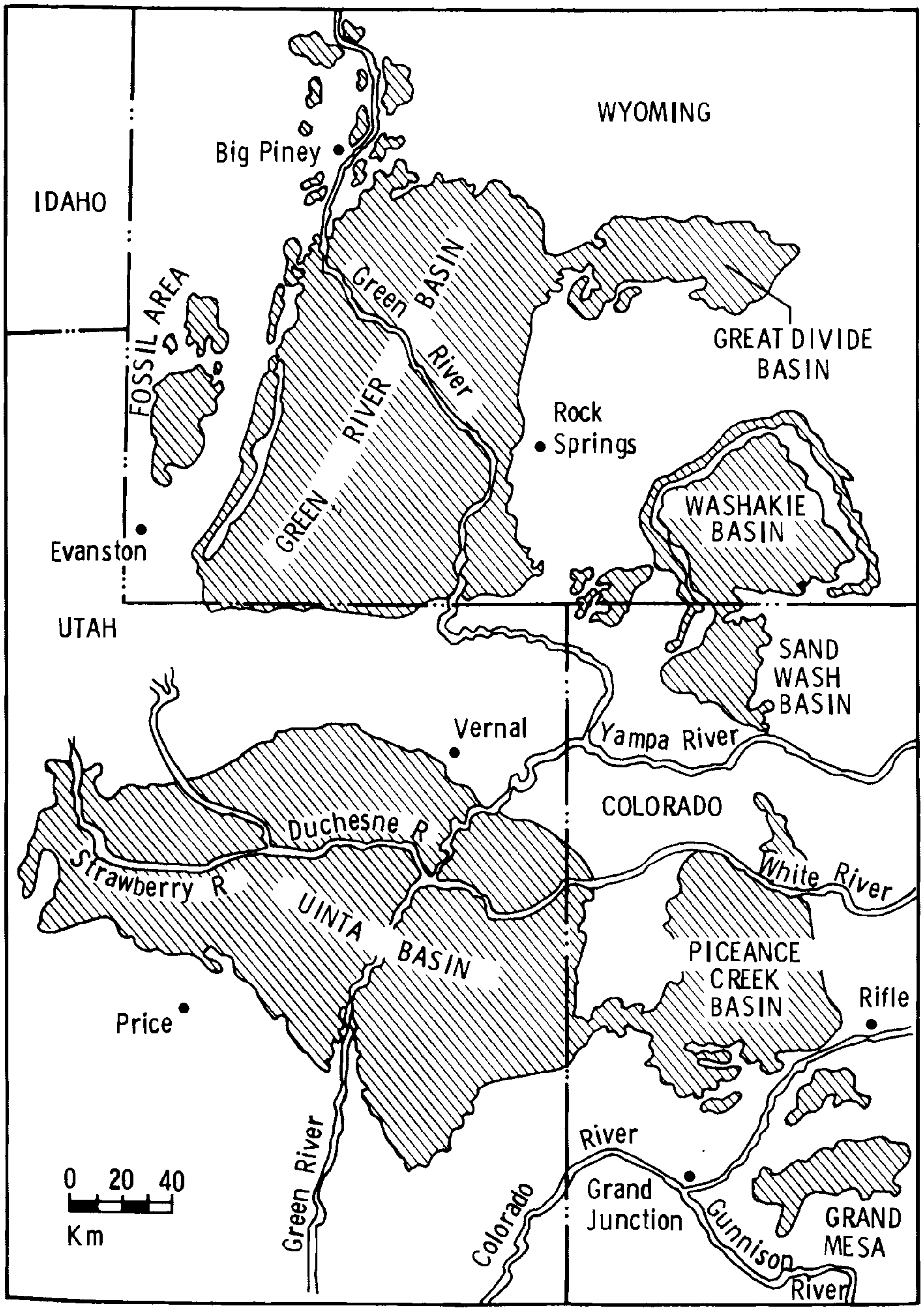


FIGURE 5.1. Distribution of Oil Shale Deposits in the Green River Formation of Colorado, Utah, and Wyoming

TABLE 5.1. A Summary of Oil Shale Activity

Development	Developer	Location	Proposed Retort	Operating Stage	Environmental Program
Federal Lease Tracts					
C-a	Gulf Standard Oil Co.	Colorado	In-situ	Shaft sinking	Baseline 1976, continuing
C-b	Occidental Oil DOE	Colorado	In-situ	Shaft sinking	Baseline 1976, continuing
U-a/U-b (White River)	Sunco Energy Develop. Co. Phillips/Sohio Nat. Res.	Utah	Surface	No development	Baseline 1976 continuing
Colony Develop. Corp.	Arco/Tosco Corp.	Colorado	Surface	Pilot Plant	DEIS 1975
Naval Oil Shale Reserve (Paraho)	Develop. Engineering Inc.	Colorado	Surface	Pilot Plant until 1978	EIS 1977
Sand Wash	Tosco Corp.	Utah	Surface	Mining	Initiated
Geokinetics	Geokinetics, Inc.	Utah	In-situ	Pilot stage	Data being taken
Equity-BX	Equity Oil Co./DOE	Colorado	In-situ	Initial drilling & fracturing	Initiated
Talley	Talley Energy Systems/DOE	Wyoming	In-situ	Initial drilling & fracturing terminated (1979)	Meteorological data only
Rock Springs	Sandia/LETC	Wyoming	In-situ	Initial drilling & fracturing	None
Superior	Superior Oil Co.	Colorado	Surface mineral extraction	Pilot Plant	None available
Long Ridge	Union Oil	Colorado	Surface	No development	None
Antrim	Dow Chemical Co. DOE	Michigan	In-situ gasification	Trial retorting	Initiated

generally lacking except for federal lease tracts in Colorado (C-a/C-b) and Utah (U-a/U-b), and the Colony Development Co. and Development Engineering, Inc. operations in Colorado. As indicated in Table 5.1 a number of environmental programs are in progress at a few other sites. A brief discussion is given on the topography, meteorology, and air quality in and around the oil shale federal lease sites C-a, C-b, and U-a, U-b. Some of the other sites are considered too but in less detail.

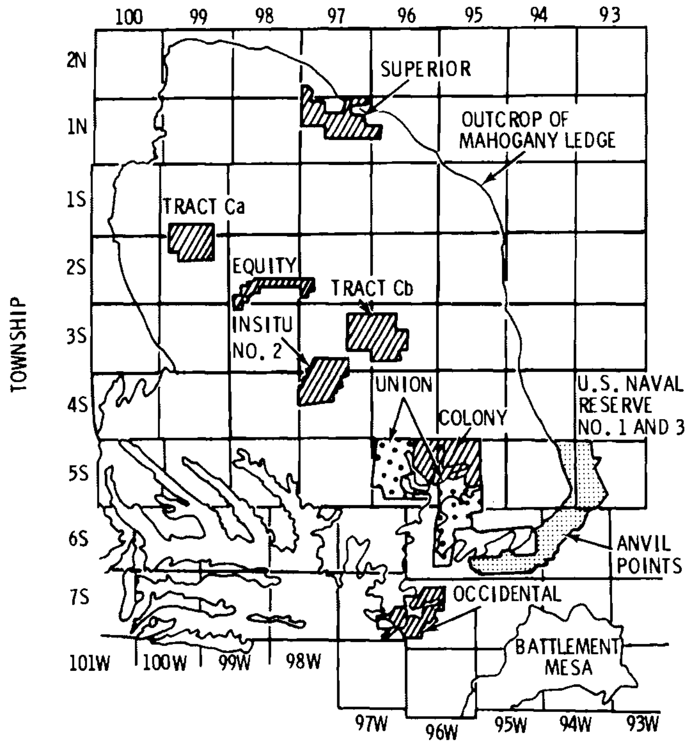


FIGURE 5.2. Locations of Potential and Existing Oil Shale Developments in the Piceance Basin, Colorado (adapted from Crawford et al. 1977)

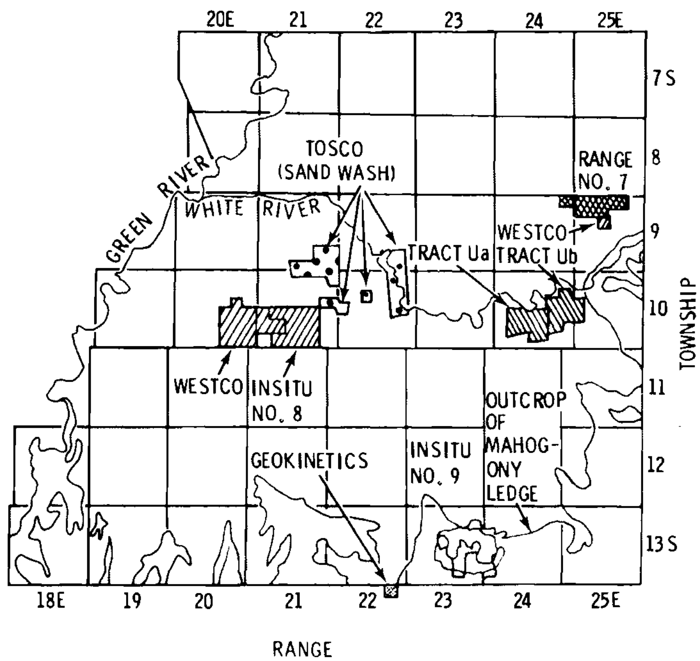


FIGURE 5.3. Locations of Potential and Existing Oil Shale Developments in the Unita Basin, Utah (adapted from Crawford et al. 1977)

5.1.1. C-a and C-b Sites

The C-a and C-b tracts are located in the Piceance Creek Basin situated between the Colorado and White Rivers on the western slope of the Rocky Mountains, at an average elevation of about 2134 m (7000 feet) (Figure 5.4). Piceance Creek Basin is an open basin with the terrain to the north generally lower than to the south, east and west. The Piceance Creek itself drains into the White River. To the northeast and east of the basin is the Flat Tops primitive area reaching to more than 3354 m (11,000 feet) in elevation, out of which the White River drains. To the south is a large high (~2600 m) massif called the Roan Plateau. To the west of the region are the Cathedral Bluffs, which are relatively steep on their west side. Their eastern side is primarily the basin itself, characterized by a series of shallow north-tonortheast draining canyons and gullies. The Piceance Creek drainage originates in the Eastern plateau region (2440 m), flows northwesterly and then turns to the north to drain into the White River.

The location of C-a and C-b tracts in the Piceance Creek Basin is shown in Figure 5.5. Tract C-a is situated near Yellow Creek on the west side of the basin about 10 km from Cathedral Bluffs and tract C-b is 2 to 3 km from the Piceance Creek and around 20 km north of the Roan Plateau. Both sites are on sloping terrain bisected by several small canyons or gullies.

Baseline meteorological and air quality studies for C-a and C-b tracts have been carried out to satisfy the provisions of the federal shale oil lease agreement (Gulf Oil Corporation--Standard Oil of Indiana 1977 and Ashland Oil--Occidental Shale, Inc. 1976). Parker (1979), Slinn (1978); and Slinn, Wolf, and Hennessey, Jr. (1978) have reviewed and critiqued the meteorological and air quality baseline studies. A number of other reports have addressed the meteorological and air quality aspects of C-a and C-b tracts such as Nevens and Rohrman (1966), EPA (1974), Kirkpatrick (1974), Meyer and Nelson (1975), Nelson (1975), Fosdick and Legatski (1977), and Fosdick et al. (1977).

The climate of the basin is similar to that of semi-arid steppe regions. Annual precipitation in the area is about 254 mm (10 inches) to 381 mm (15 inches) with over half of the moisture falling as snow. Stagnant high pressure systems often persist for several days resulting in a high frequency

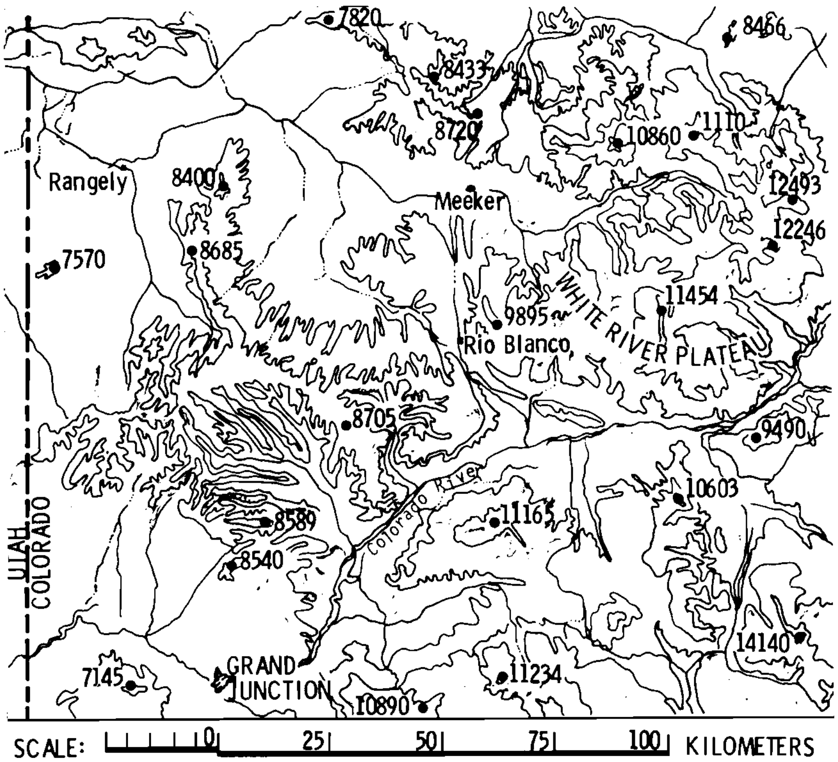


FIGURE 5.4. General Topography of the Piceance Creek Basin

of clear sunny days with light winds and large diurnal temperature changes. Meteorological data indicate a high degree of temporal and spatial variability in this area as might be expected for rough terrain.

The prevailing geostrophic winds are generally from west to southwest and exist year-round, except for temporary interruptions because of migratory frontal or cold low systems. Winds at the top of the higher terrain such as Cathedral Bluffs and Roan Plateau are strongly influenced by the upper-level geostrophic winds. Because of the orientation of this higher terrain with respect to the prevailing wind, a shielding effect exists for most of the basin. This is enhanced by the overall gentle downslope of the basin to the north and northeast.

The combination of topography and geostrophic flow influences the winds over the basin regarding dispersion potential in three ways:

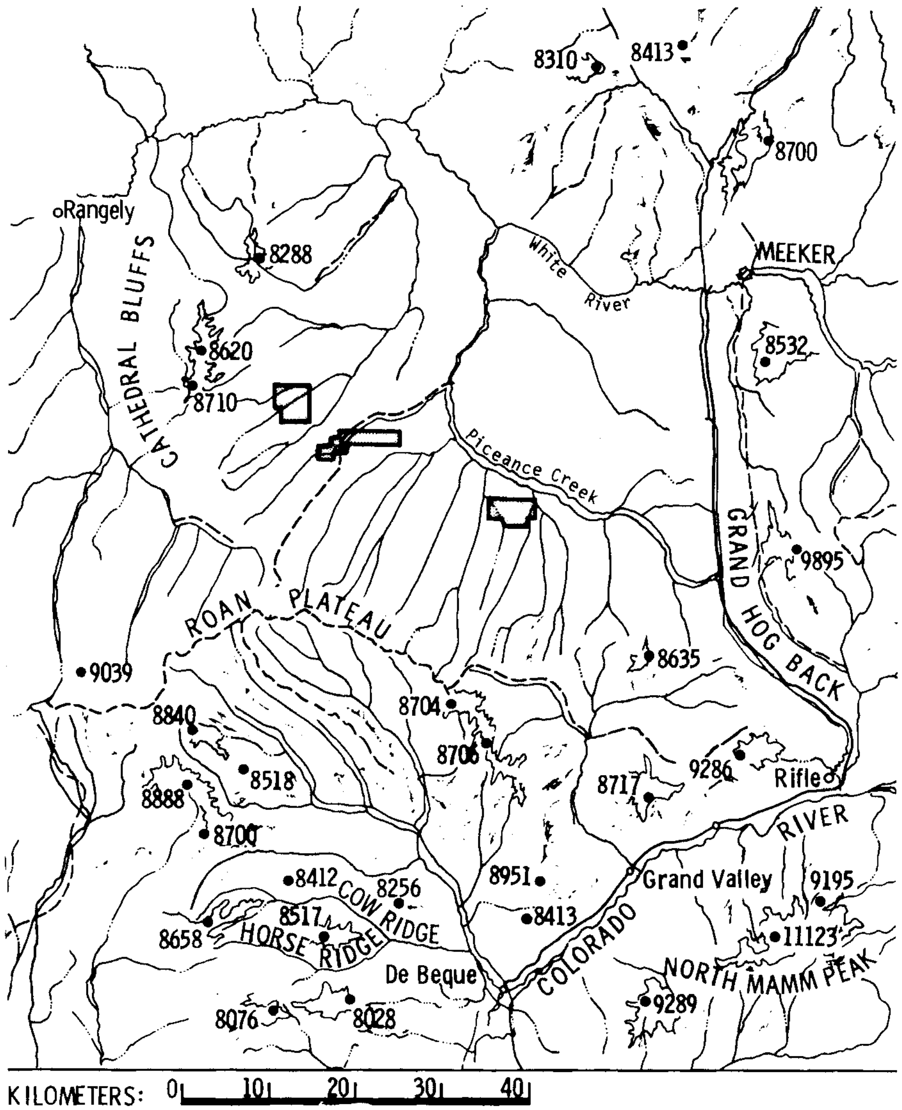


FIGURE 5.5. Location of Federal Lease Tracts C-a and C-b and the Equity Development Site in the Piceance Creek Basin

1. in the enhancement of mechanical-generated turbulence (separation) along the top and to the lee of the plateau and bluff as the wind is forced up and over the summits.

2. in the downslope of the basin with increasing distance. This, in effect, should result in the near-ridge-level winds decoupling from the surface winds within the basin especially during inversion conditions.

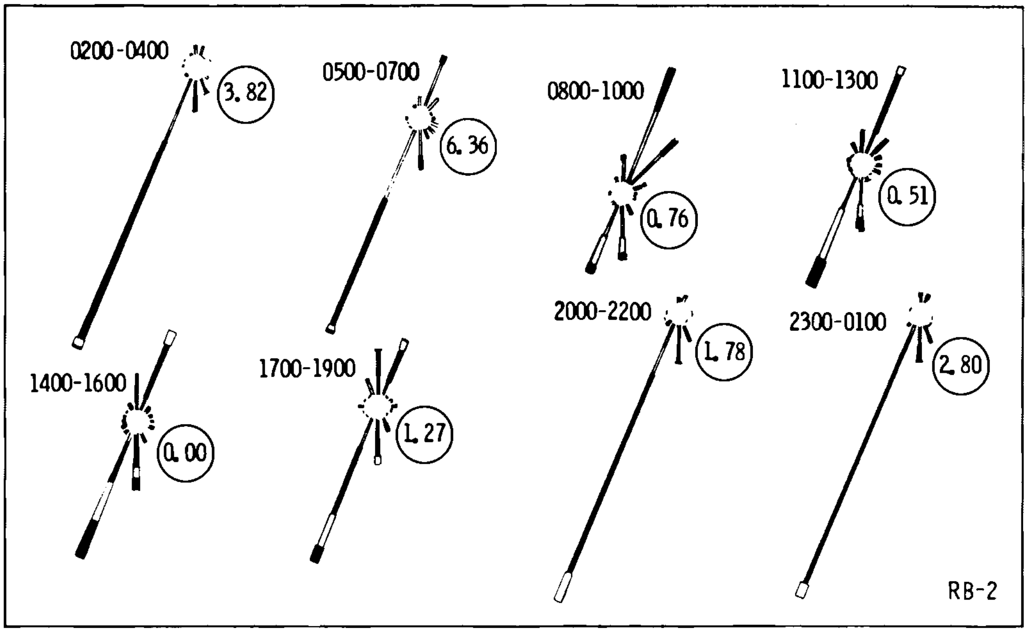
3. the development of local wind systems.

As might be expected, wind data from various locations around the C-a and C-b sites show a definite slope-valley wind component (Figure 5.6). Gullies and canyons greater than 61-m (200-ft) deep may restrict the movement of air to upslope and downslope (Meyer and Nelson 1975). Winds observed on the basin slopes indicate a high frequency of nocturnal downslope or drainage winds.

Temperature data from Meyer and Nelson (1975), Nelson (1975) and Ashland Oil--Occidental Oil Shale, Inc. (1976) indicate that inversion formation and dissipation within the basin follow extremely predictable diurnal patterns. The temperature structure over the tracts follow the dry adiabatic lapse rate well above 300 m on clear days without inversions and a wet adiabatic behavior on moist days with very few surface-based stable layers associated with precipitation. The air from the surface to the 300-m level exhibits a complex profile with inversions in the morning hours that dissipate by early afternoon or late morning (Figure 5.7). Temperature inversions were pronounced in the valley below the tracts. Typical statistics compiled over four seasons of upper air data are shown in Table 5.2.

The formation and dissipation of temperature inversions so well defined could lead to serious pollution problems if large pollutant releases were made within the lower basins. Trapping and fumigation could occur if plumes are released on the lower slopes of the basin. Overall, the dispersion potential of the basin is good when source releases are made into the air above the plateau regions and could be poor for releases made within the basin, although trapping inversions seldom seem to exist throughout a 24-hr period.

The Colorado oil shale area is currently designated as Class II under Prevention of Significant Deterioration (PSD) regulations and as Category I under State Ambient Air Quality Standards (SAAGS) (Table 5.3). Ambient air quality was monitored in accordance with the lease agreement and to satisfy



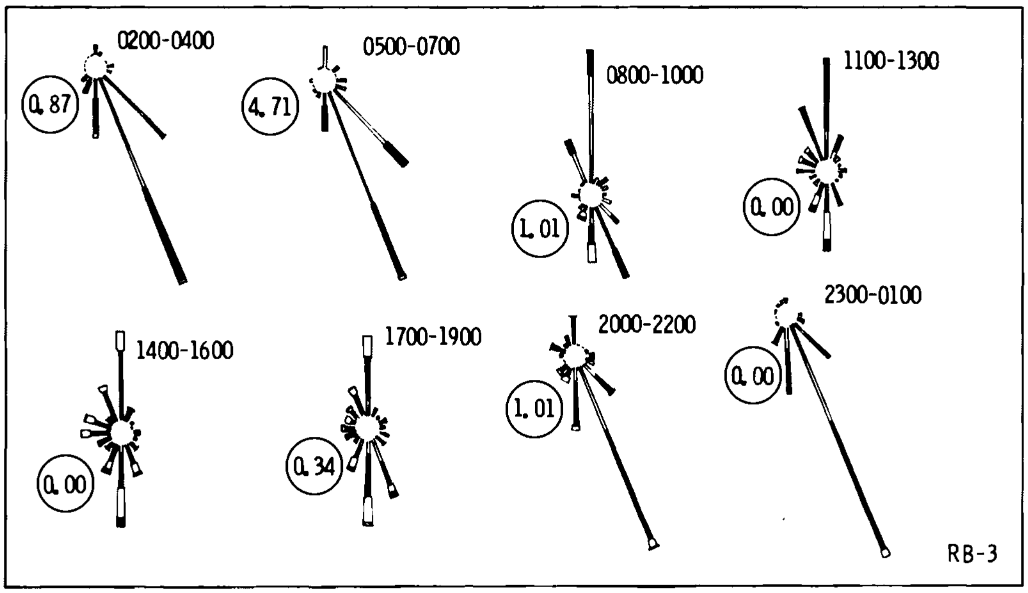


FIGURE 5.6. Wind Rose Plots of Wind Data Measured at Rio Blanco Stations RB-2 and RB-3 (adapted after Nelson 1975)

federal and State of Colorado regulations. Table 5.4 summarizes the baseline air quality data. Mean pollution concentrations reviewed were low compared to the federal standards. However, particulate, ozone and non-methane hydrocarbon occasionally exceeded the federal standards. The particulate loading has been

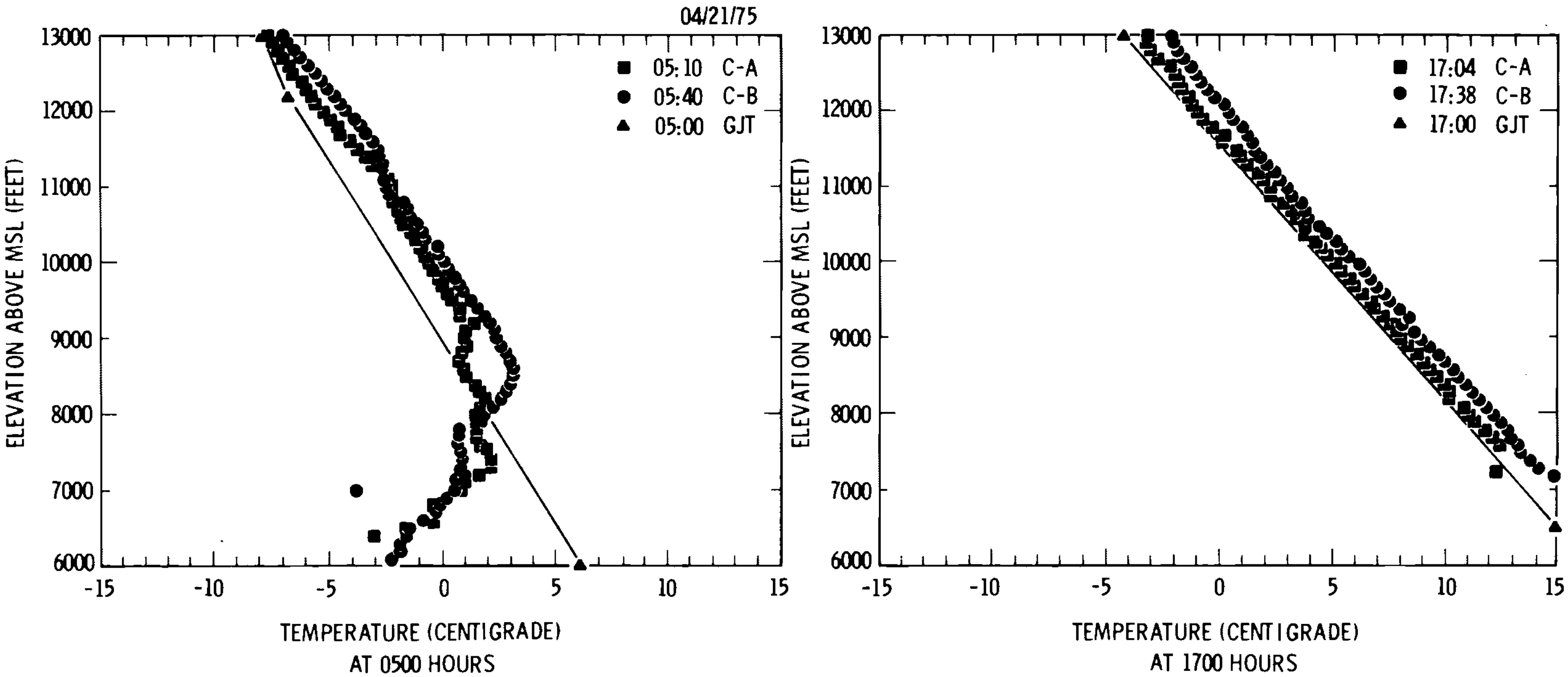


FIGURE 5.7. Air Temperauture Soundings on April 21, 1975, Over Federal Oil Shale Tracts C-a, C-b, and Grand Junction (adapted from Meyer and Nelson 1975)

TABLE 5.2. Inversion Statistics Over the Piceance Creek Basin

	Frequency of Inversion, %	Frequency of Inversion Plateau, %	Mean Height of Inversion Existing Above Plateau (feet above plateau)	Mean Lapse Rate Within Inversions Below Plateau (°C/100m)	Mean Lapse Rate Within Inversions Above Plateau (°C/100m)
Fall 74	100	50	750	2.38	1.60
Winter 75	60	27	725	3.86	0.42
Spring 75	47	20	450	1.84	0.69
Summer 75	100	53	1,000	2.99	0.44

TABLE 5.3a. National Ambient Air Quality Standards (NAAQS)

Emissions	Averaging Time	Primary Standards for Human Health (μg³/m)	Secondary Standards for Human Health (μg³/m)
Sulfur Oxides (SO_2)	Annual	80	None
	24-hr[a]	365	None
	3-hr[a]	None	1,800
Nitrogen Dioxide (NO_2)	Annual	100	100
Particulates	Annual[b]	75	60
	24-hr[a]	260	150
Hydrocarbons (corrected for CH_4)	3-hr[a] (6 to 9 AM)	160	160
Carbon Monoxide	8-hr[a]	10,000	10,000
	1-hr[a]	40,000	40,000
Oxidant (corrected for NO_2, and SO_2)	1-hr[a]	160	160

(a) Not to be exceeded more than once per year.
(b) Geometric mean of the 24-hr concentration.

TABLE 5.3b. Maximum Allowable Federal Increments for Area Classes Prevention of Significant Deterioration Regulations (PSD)

Emission	Averaging	Class I ($\mu g/m^3$)	Class II ($\mu g/m^3$)	Class III ($\mu g/m^3$)
Particulates	Annual[a]	5	10	Same as NAAQS
	24-hr max	10	30	
Sulfur Dioxide	Annual[b]	2	15	
	24-hr max	5	100	
	3-hr max	25	700	

(a) Geometric mean of the 24-hr concentrations.
(b) Arithmetic mean of the 24-hr concentrations.

TABLE 5.3c. State of Colorado Ambient Air Quality Standards (SAAQS)

Emission	Time	Maximum Allowable Increments Over Baseline ($\mu g/m$) Category I	Category II	Maximum Allowable Concentrations ($\mu g/m$) Category III
SO_2	Annual	3	15	60
	24-hr max	15	100	260
	3-hr max	75	700	1,300
Particulates[a]	Annual[b]	45		
	24-hr max[c]	150		

(a) Maximum allowable ambient air concentrations in non-designated areas.
(b) Arithmetic mean of the 24-hr concentrations.
(c) Not to be exceeded more than once in a 12-month period.

ascribed to natural fugitive dust. The hydrocarbon loading has been ascribed to natural vegetation and volatilization of compounds from surface shale. The ozone loading has been ascribed to long-distance transport from urban sources or from stratospheric injection of ozone associated with passages of storms. Since the air quality is generally good, the visual range within the basin is exceptional, often being in excess of 160 km (100 miles).

TABLE 5.4. Summary of Air Quality Data from Tract C-a and C-b Baseline Study

Criteria Pollutant	Tract C-a	TRACT C-b
Particulate		
Maximum 24 hr ($\mu g/m^3$)	469	178
Annual Mean ($\mu g/m^3$)	18	11
No. occurences exceeding NAAQS[a]	5	4
SO_2		
Maximum 3 hr ($\mu g/m^3$)	345	88
Maximum 24 hr ($\mu g/m^3$)	82	112
Annual mean ($\mu g/m^3$)	11	1
NO_x		
Annual Mean ($\mu g/m^3$)	4	21
CO		
Maximum 1 hr ($\mu g/m^3$)	5,823	2,597
Maximum 8 hr ($\mu g/m^3$)	4,825	4,502
NMHC[b]		
Maximum 3 hr ($\mu g/m^3$)	505	2,597
No. occurrences exceeding NAAQS[a]	94	262
Oxidant		
Maximum 1 hr ($\mu g/m^3$)	177	160
No. occurrences exceeding NAAQS[a]	5	0

(a) National ambient air quality standards.
(b) Non-methane hydrocarbons.

5.1.2 U-a and U-b Sites

Tracts U-a and U-b are within the eastern portion of the Uinta Basin, a broad, northwest-facing asymmetric basin on the northeastern edge of the Colorado Plateau (Figure 5.8). Major land features bordering the basin are the Uinta Mountains on the north, the Wasatch Range on the west, the Roan Cliffs on the south, and the highlands associated with the subsurface Douglas Creek Arch along the eastern edge.

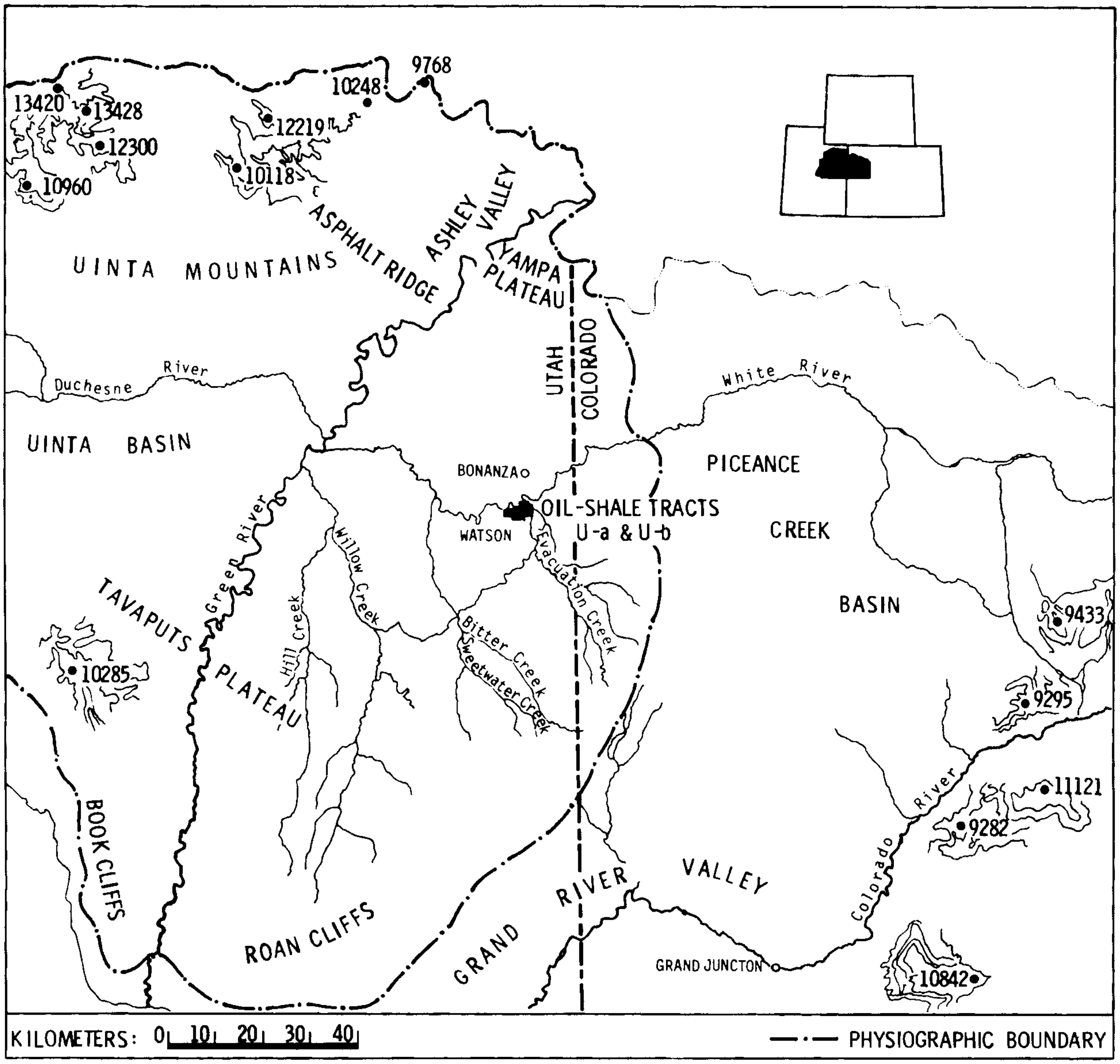


FIGURE 5.8. Physiographic Map of the Unita Basin Region (Utah) and the Location of Federal Oil Shale Lease Tracts U-a and U-b

Tracts U-a and U-b are bordered by the White River to the north, Hells Hole Canyon to the east, upland areas of elevations between 1829 m and 1890 m (6000 ft and 6200 ft) to the south, and Asphalt Wash to the west. The region near the Roan Cliffs and westward to the Green River is a gently north-sloping, highly dissected plateau characterized by steep-walled canyons with ephemeral streams or dry washes. The elevations on tracts U-a and U-b range from 1500 m to 1890 m (4920 ft to 6200 ft) for a maximum relief of 390 m (1280 ft), as seen on Figure 5.9. In general, the lands within tracts U-a and U-b slope toward the east-west channel of the White River.

Within the tracts, the landscape is composed of a series of north-south valleys separated by narrow, elongated mesas. From east to west, three dominant ridge sections are just west of north, north, just east of north, respectively. The maximum dimensions of these landforms are oriented perpendicular to the direction of the White River. Geologic structure and erosion have influenced the development of several landforms on tracts U-a and U-b, especially rock pinnacles and balanced rocks. These numerous and prominent forms on or near ridgetops have developed by progressive headward erosion by drainages along vertical fractures or joints. In addition, large rock faces on canyon walls along the White River show evidence of massive rock failures. These have occurred along vertical fractures that are generally perpendicular to the bedding.

The White River Oil Shale Project plans to develop tracts U-a and U-b in Uinta County, Utah. A final environmntal baseline report was issued concerning monitoring from October 1974 through January 1977 (VTN Colorado, Inc. 1977). The Air Resource section of the report was done by Aero-Viroment, Inc. who evaluated existing meteorological conditions and air quality on the tracts.

The prevailing westerly air currents that reach the tracts are comparatively dry having lost their moisture to the higher terrain to the west (Wasatch Mountains), northwest (Uinta Mountains) and south (Tavaputs Plateau). Clear skies prevail most of the year, with strong insolation during the day and rapid nocturnal cooling during the night, resulting in fairly wide daily ranges in temperatures.

High-pressure areas that are most prevalent in the winter tend to settle over the intermountain area for a week or more. These stagnating high-pressure cells are usually associated with extended periods of light winds, generally 1 to 2 ms^{-1}. Occasional storms with high winds ($\geq$10 ms^{-1}) are usually associated with the passage of migratory low pressure and frontal systems through the area. The intensity and frequency of these weather systems are highest during February through April.

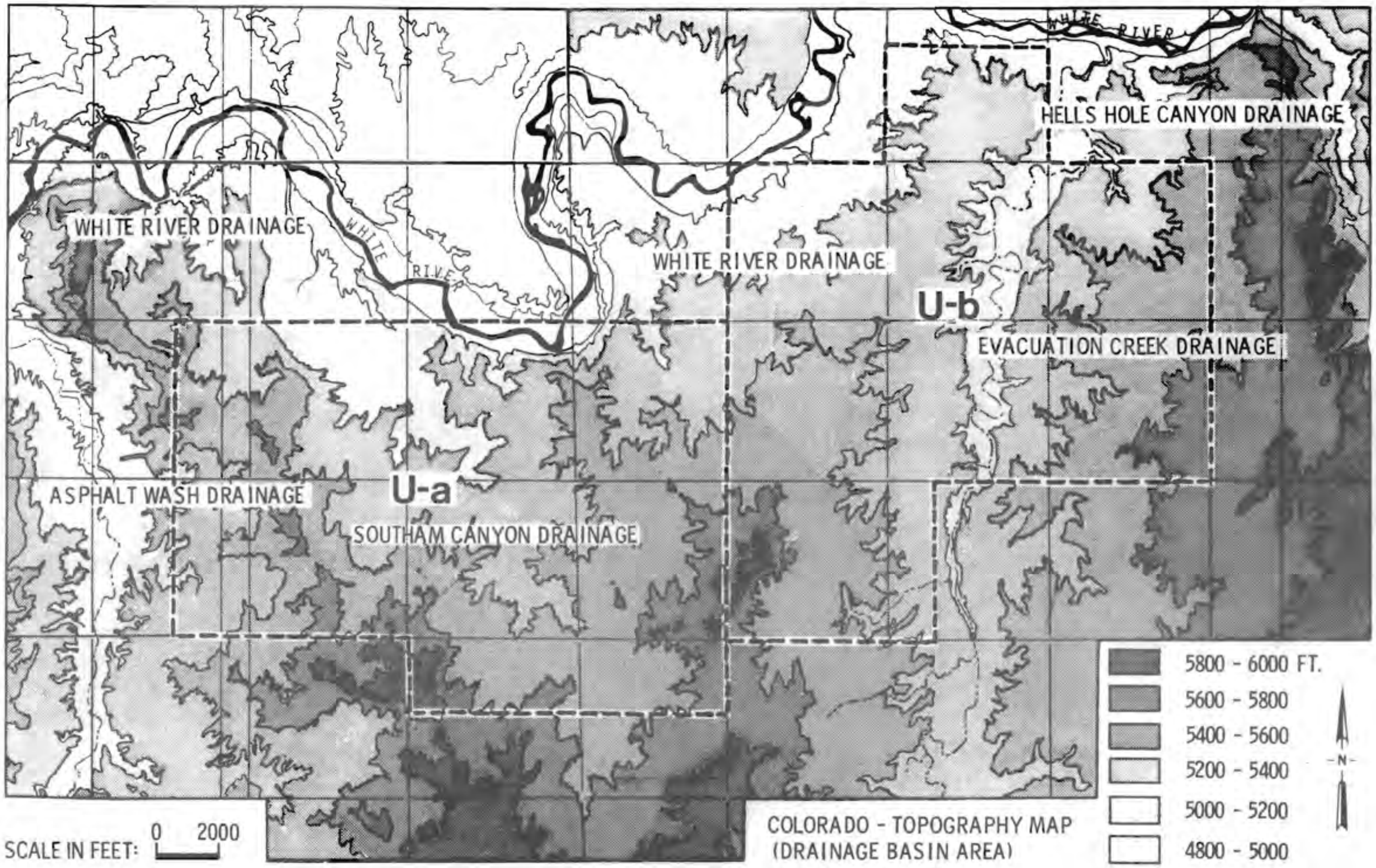


FIGURE 5.9. Topography of the Federal Oil Shale Lease Tracts U-a and U-b

Winds above the first kilometer are usually from the west with an average speed of about 8 ms^{-1} during winter and spring and about 4 ms^{-1} in summer and fall. Near the surface, excluding days with storms, the winds follow a diurnal cycle being lower at night and higher in the afternoon. The surface airflow pattern on the tracts is complicated by the effects of the rugged terrain. Generally, drainage and downslope conditions dominate in the early morning hours with airflow toward low terrain and down the White River. In the afternoon, the winds are usually out of the west or southwest. Wind speeds over the ridges and widely exposed terrain are higher than those protected in canyons and gullies (Figure 5.10).

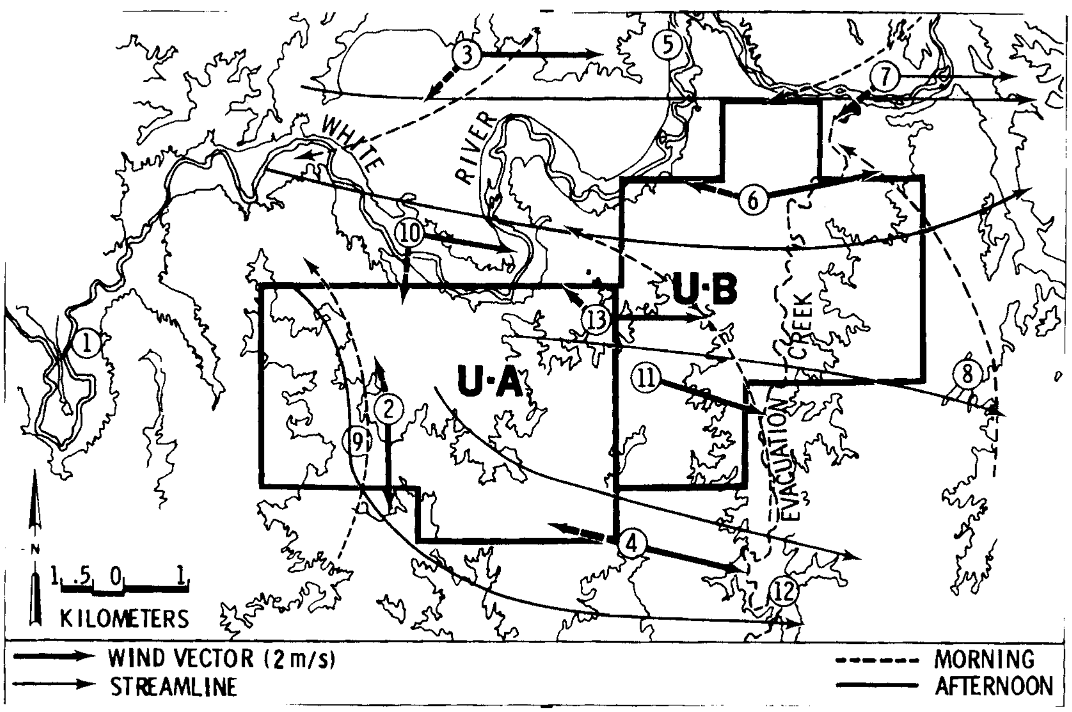


FIGURE 5.10. Typical Airflow Patterns on Tracts U-a and U-b in the Morning and Afternoon during October 1976

Surface-based temperature inversions usually occur in the morning and dissipate in the afternoon. The average thickness of the morning inversion increases from about 243 m in winter to 253 m in spring and from 362 m in summer to 209 m in the fall. The strength of the inversion in winter, spring, and fall is about 2°C/100 m and 1°C/100 m in summer.

The dispersion of plumes is related to the product of σ_v and σ_w where σ_v and σ_w are the root mean-square turbulence fluctuation in the lateral wind speed (v) and the vertical wind speed (w) (MacCready, Baboolal, and Lissaman 1974). These were measured at the tract sites. The diffusivity is typically between 0.2 to 4.0 ms^{-1}. The frequency of occurrence of high diffusivity (>1.0 ms^{-1}) is greater in spring and summer, and the frequency of occurrence of low diffusivity (<0.2 ms^{-1}) is greatest in the winter. Sheltered locations have a high frequency of low diffusivity while exposed sites have higher frequencies of high diffusivity.

The most frequently occurring atmospheric stability conditions on an annual basis are stability classes D and E (neutral and slightly stable). This combined with the preceding information on diffusivity indicates that dispersion of pollutants will be of greatest concern in winter, especially during the early morning.

The air quality on the tracts is representative of a naturally occurring background air environment because of the lack of significant anthropogenic sources of air contaminants in the area. A summary of the air quality data is given in Table 5.5. Ozone and non-methane hydrocarbons (NMHC) have exceeded the NAAQS on several occasions but have been tentatively explained by the natural environment, atmospheric processes, and long-range transport.

Particulate concentrations are typical of rural areas, where dusts are generated primarily from wind and vehicles traveling on unpaved roads. Particulate concentrations are lower when the ground is covered with snow. There are presently no concentrations exceeding any suspended particulate standards. Trace metals present in the air are mostly of soil origin, indicating the relative absence of anthropogenic activities. The low particulate concentrations in air result in exceptional visibility as indicated in Table 5.6.

TABLE 5.5. Summary of Air Quality Data from Tracts U-a/U-b Baseline Study

Criteria Pollutant		Tracts U-a/U-b
Particulate		
Maximum 24 hrs	($\mu g/m^3$)	127
Annual Mean	($\mu g/m^3$)	22
SO_2		
Maximum 3 hrs	($\mu g/m^3$)	40
Maximum 24 hrs	($\mu g/m^3$)	25
Annual Mean	($\mu g/m^3$)	10
NO_x		
Annual mean	($\mu g/m^3$)	10
CO		
Maximum 1 hr	($\mu g/m^3$)	5,200
Maximum 8 hrs	($\mu g/m^3$)	3,900
NMHC[(a)]		
Maximum 3 hrs	($\mu g/m^3$)	1,970
No. of occurrences exceeding NAAQS*		138
Oxidant		
Maximum 1 hrs	($\mu g/m^3$)	190
No. occurrences exceeding NAAQS[(b)]		42

(a) Non-methane hydrocarbons.
(b) National Ambient Air Quality Standards.

TABLE 5.6. Local Visual Range Values (in km) on the Tracts in 1976, km

	Maximum	Minimum	Average	No. of Observations
Winter	235	46	122	2184
Spring	235	21	145	2205
Summer	235	67	131	2206
Fall	235	50	126	2184

5.1.3. Other Sites

As indicated in Table 5.1 and Figures 5.2 and 5.3 a number of other oil shale operations have taken place during the past few years. These include sites in Colorado, Utah and Wyoming.

5.1.3.1. Piceance Basin

Equity Oil Company, in cooperation with DOE, is operating an in-situ project about 10 km (6 miles) south of the federal lease tract C-a in the Piceance Creek Basin. An environmental research plan is underway. However, the topography, meteorology, and air quality should be similar to that of the C-a and C-b sites (Dougan 1978).

Superior Oil Company has a private venture for a multi-mineral process for recovering oil, alumina, nahcolite and soda ash from oil shale at their site in the Piceance Creek Basin. Their site is just south of White River about 35 km (27 miles) west of Meekers.

Colony Development Operation has submitted an environmental impact assessment for construction and operation of a shale oil production plant at a site in the Roan Plateau in the vicinity of Parachute Creek, north of Grand Valley (Colony Development Operation Vol I and II, 1975). A semi-works retort has been built and operated near the head of Parachute Creek. Battelle, Pacific Northwest Laboratories conducted three air quality impact studies for the Parachute Creek Valley and Roan Plateau (Battelle 1972 and 1973, and Bander and Wolf 1975). The climate and meteorology of the area are similar to the Piceance Basin but the topography is very rugged consisting of multiple canyons and steep cliffs. Parachute Creek drainage enters the Colorado River Valley near Grand Valley. Studies have indicated that local wind systems are well developed, especially the nocturnal drainage wind.

The Naval Oil Shale Reserve Operations at Anvil Point under Development Engineering, Inc., have had sporadic operations for several years. By statute, the plant can only be used for research, development, testing, evaluation and demonstration work (Thorne 1952). The plant is located approximately 15 km west of the town of Rifle beneath the Roan Cliffs in the Colorado River valley.

A limited baseline study was done by Development Engineering, Inc. (1976) but comprehensive air quality and meteorological studies have not been made to date.

Occidental Oil Shale, Inc., has been operating a modified in-situ operation since 1972 at Logan Wash between DeBeque and Grand Valley (Jackson 1976). Work on a fullscale in-situ retort has shifted to lease tract C-b. Only limited air quality and meteorology studies have been made to date. Topography is similar in nature to the Anvil Point location.

5.1.3.2 Uinta Basin

Geokinetics in cooperation with DOE is conducting an in-situ field operation at a location some 112 km (70 miles) South of Vernal, Utah, near the Tavaputs Plateau. The site is at an elevation of about 1950 km on a gradual north-south slope. The terrain is less rugged than the U-a and U-b sites. An environmental baseline study on air quality and meteorology is underway.

TOSCO corporation is in the development stages of their Sand Wash Project, which is located approximately 24 km (15 miles) west of the federal lease tracts U-a and U-b.

5.1.3.3 Green River Basin

Talley Energy Systems and Sandia Laboratoties in cooperation with DOE-Laramie Energy Technology Center have conducted in-situ field operations at a site approximately 11 km (7 miles) west of the town of Rock Springs, Wyoming. The site is between the Green River and a 2380-m ridge. No established program for baseline date has been initiated to date.

5.1.4. Summary

Generally, the oil shale development sites are located in at least three topographically distinct areas: the Green River Basin, the Uinta Basin and the Piceance Basin. Topographically, the Piceance Basin in Colorado has the roughest terrain of the three. The Green River Basin has the least complex terrain. Examination of characteristic plan-profile types in Appendix I Figure I.3 suggests types 1 and 2 (40% to 60% of area) for the Piceance Basin, types 3 and 6 (<40% of area) for the Green River basin and types 2 and 3 (40% to 60% of area) for the Uinta Basin.

In respect to climate and meteorology, the Piceance Basin probably has a higher incidence of slope-valley (local) winds because of numerous canyons and valleys. The Piceance and Uinta basins have a high incidence of temperature inversions probably relatively higher than the Green River Basin. High winds are generally more frequent in the Green River Basin because of the terrain and storm tracks; thus, temperature inversions and slope-valley winds systems are relatively less frequent there than in the other basins. Uinta basin probably has a higher frequency of strong winds than the Piceance basin.

At the present time, the air quality in all three basins is excellent most of the time because of the lack of anthropogenic activity. Some natural causes such as dust storms or vegetation emissions, result in temporary degradation in the air quality. Long-range transport of anthropogenic pollutants could become more of a problem in the future.

5.2 GEOTHERMAL RESOURCE SITES

Geothermal resources are derived from the distribution of temperatures and thermal energy beneath the earth's surface. This thermal energy, whether dominated by vapor or by hot water, is generally associated with tectonic-plate boundaries and volcanic activity. Temperatures below the earth's surface are controlled principally by conductive flow of heat through solid rocks by convective flow in circulating fluids, or by mass transfer in magma. Geothermal systems are generally divided into three types: 1) those associated with conductive thermal gradiants, 2) hydrothermal-convective systems, and 3) geopressure systems. Under the hydrothermal-convective system exist the two subtypes: hot-water and vapor-dominated (dry-steam) resources. Both the hot-water and vapor-dominated systems have further sub-types (White 1973).

The geothermal resources of primary interest are those that may become economically competitive within the next 25 years. This restriction requires focusing on hydrothermal resources, hot-water or vapor-dominated, which have the highest temperatures (>170°C). Since vapor-dominated systems are very limited in the United States (The Geysers),[a] much of the present geothermal development is with hot-water systems in the western states.

The National Geophysical and Solar-Terrestrial Data Center, in cooperation with the ERDA (DOE) and the USGS, produced a multi-colored data map depicting the known geothermal energy resources (KGRA) of the western United States excluding Alaska and Hawaii. Most of the data were taken from existing data compilations and show those areas where geothermal resources are known or suspected (e.g., locations of hot springs, locations of volcanos and volcanic cones of relatively recent age and geothermally classified lands) as well as the areas where there are no indications of geothermal resources (Grim, Clark, and Morris 1977, and Grim 1978). Appendix II lists the known geothermal resources areas by state as identified by Grim and others. Tentative terrain types are also included with the KGRAs in Appendix II.

The geothermal industry in the United States has been growing rapidly and undergoing significant changes from its early focus on the Geysers dry-stream field in California. Present geothermal developments, i.e., well drilling, pilot and commercial projects, are generally confined to the states of California, Idaho, Nevada, New Mexico, Utah and Hawaii. In this section the various developmental geothermal resources sites in the western states are reviewed primarily with respect to their topographic setting and dispersion climate.

5.2.1. The Geysers--Calistoga KGRA

Our nation's major geothermal development is in the Geysers-Calistoga KGRA that encompasses four Californian counties: Lake, Sonoma, Mendocino, and Napa (see Figure 5.11). The Geysers site is located about 40 km north of the city of Santa Rosa and around 60 to 70 km east of the Pacific Ocean. Thirteen units are now in operation at The Geysers, providing 743 MW of power, and several additional power plants are under construction. By 1983, it is expected that eighteen units will generate 1248 MW of electricity.

(a) A second vapor-dominated system appears to have been discovered at the Dixie Valley KGRA in Nevada.

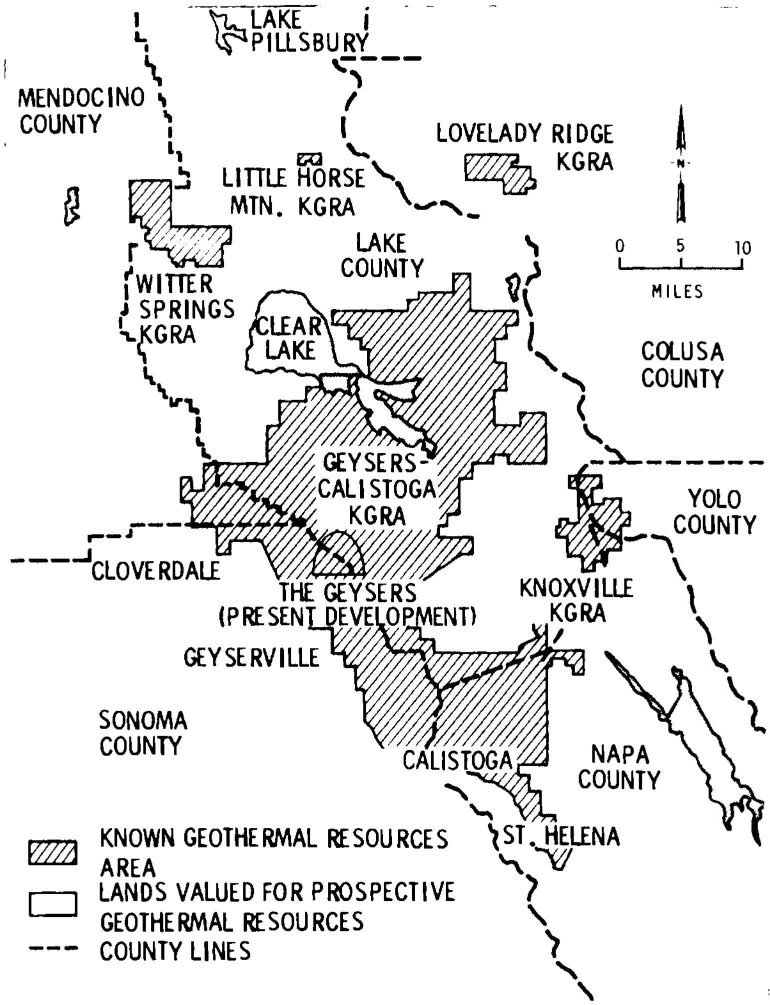


FIGURE 5.11. The Areal Extent of the Geysers-Calistoga KGRA. The area includes six California counties: 1) Lake, 2) Sonora, 3) Mendocino, 4) Napa, 5) Yolo, and 6) Colusa

The topography of the Geysers area is shown in Figure 5.12. The Mayacmas mountains, a major ridgeline, runs nearly north and south, forming the border between Lake and Sonoma counties. This ridge separates a populated resort area to the east from an unpopulated area immediately to the west. A series of small valleys run northwesterly and southeasterly from the ridgeline. Some of these valleys measure 4 km from ridge to ridge and are around 600 m deep. The power plants on the west side vary in altitude from about 420 m to 960 m. To the north, the Geyser KGRA is bordered by orchards, farm lands and a lake.

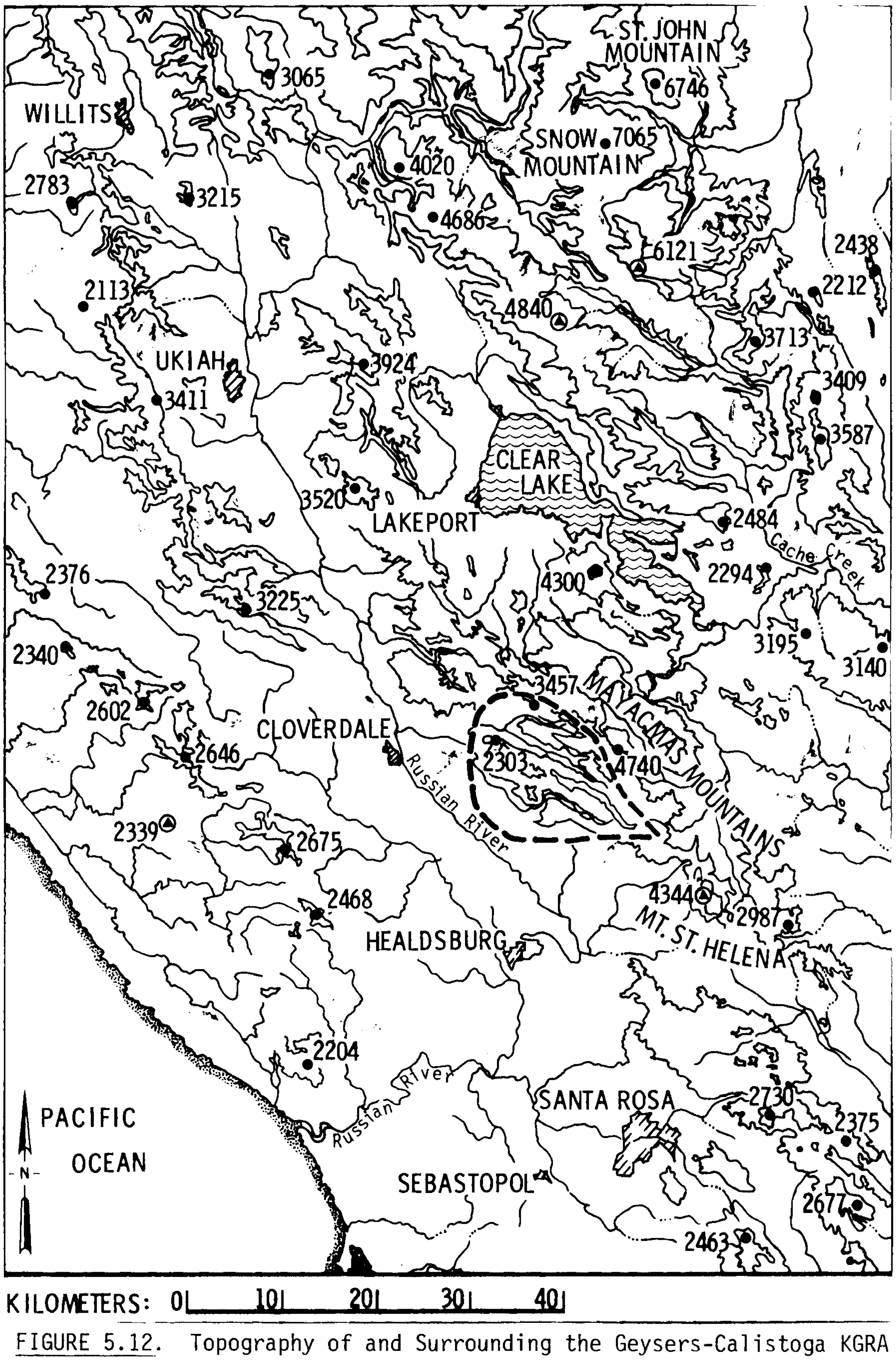


FIGURE 5.12. Topography of and Surrounding the Geysers-Calistoga KGRA

New development of wells and power plants are planned to the east in Anderson Creek Valley. Approximate dimensions of the valley are:

Width of valley floor	1 to 2 km
Distance between ridge lines	~8 km
Length of valley	~10 km
Height of ridge lines above valley floor (average)	~731 m
Slope of valley	~0.041 to 0.224

Highest mountain peaks in the area are Mt. Saint Helena (1303 m), Cobb Mtn. (1422 m), and Mt. Konocti (1290 m).

One of the most critical issues at the Geysers-Calistoga KGRA is the control of hydrogen sulfide emissions. Hydrogen sulfide is a component of the geothermal fluid and is released to the atmosphere at several locations within the power plant complex. In the atmosphere, hydrogen sulfide is transported to surrounding regions and can result in an odor nuisance if the concentration is sufficiently high. After hydrogen sulfide, boron emissions are of primary concern. Of secondary importance are emissions of ammonia, mercury, arsenic, and sulfates.

A network of stations for monitoring H_2S, temperature and wind have provided baseline data on hydrogen sulfide concentrations in the area (Cavanagh and Ruff 1979) and the transport and diffusion of H_2S in the Geysers area (Ruff 1979 and Steffen et al. 1978). Field tracer experiments using SF_6 and $CBrF_3$ have been conducted (MRI 1978, and Knuth and Jensen 1978). Air quality modeling with computer codes has been performed by ASCOT (1980a and b) and others (Rosen and Molenkamp, 1978). The Geothermal Overview Project is identifying, summarizing, and assessing the environmetal issues of The Geysers-Calistoga KGRA (Phelps et al. 1979, Ermak and Phelps 1979, and Rosen and Molenkamp 1978). ASCOT is conducting field and modeling efforts concerning the local wind systems, particularly drainage winds, in the Geysers area. Multi-laboratory field efforts were conducted in July 1979 and September 1980. Another field effort is planned for the summer of 1981.

The results to date indicate that the complex terrain in The Geysers-Calistoga KGRA induces a partial decoupling of the quasi-gradient flow near the ridge line and the local wind systems in the valleys and canyons. During the morning and mid-day, the slopes of the Mayacmas mountains are heated by insolation. Winds near the surface blow upslope towards the ridge crest and transport the emissions west of the ridge towards the crest and into Lake County. After sunset, the slopes are cooled, resulting in downslope and drainage winds in the valleys. During the summer, the monsoon-sea breeze extends into the Geysers area and results in relatively strong westerly winds during the afternoon and, at times, evening hours. These winds essentially modify or destroy the drainage wind at night. Vertical circulation cells or rotors are set up over the slopes and valleys because of the uneven heating of the valley slopes. The downdrafts of these cells are suspected of entraining H_2S into some valleys and to surface levels.

5.2.2 Imperial Valley

The Imperial Valley in southern California has six KGRAs that have potential for major geothermal development (Figure 5.13). The current activities are to develop hot geothermal brines for their potash content and the heat content for both electric and nonelectric application in the form of process heat.

The geothermal fields of the Imperial Valley are water-dominated and lie in the Salton Basin. This rift valley, called the Salton Trough, is a landward extension of the Gulf of California. The valley is around 70 m below sea level and is approximately 80-km wide narrowing to around 40 km in the north. The land is essentially flat and extensively irrigated to the south of the Salton Sea. The topography is shown in Figure 5.14. Since the principal part of the valley is from 30 to 50 km from higher terrain its setting is not typical of complex terrain.

The climate of the area has typical desert characteristics: high [average July temperature ~32°C (90°F)] temperatures, low relative humidity (~26%), clear skies, strong solar insolation, good visibility, and very little precipitation (average yearly ~69 mm). Prevailing surface winds are generally northerly from November through February, westerly from March through June, southwesterly in July and August, and westerly again in September and October. Their mean annual speed is 2.3 ms^{-1}. Typical radiation inversions are

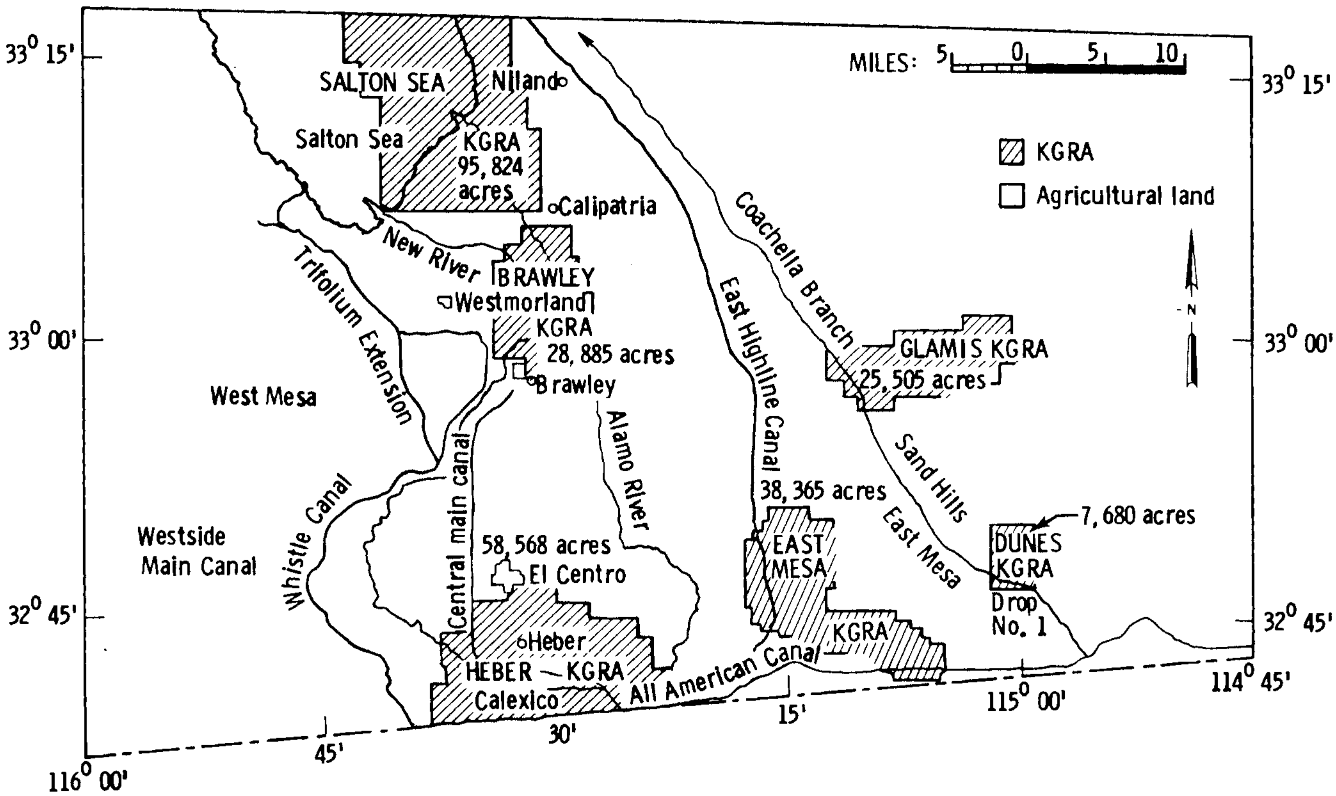


FIGURE 5.13. The Known Geothermal Resource Areas (KGRAs) of Imperial Valley, California

frequent with 90% of the days in each month showing the presence of an inversion. Average height is around 370 m. These inversions form during the early evening and break up about 1100 hours during the morning.

The assessment of environmental impacts of a potential geothermal energy development in the Imperial Valley is underway (Anspaugh and Phelps 1976, Layton and Ermak 1976, Shinn 1976, and Gudiksen et al. 1977). Some preliminary assessments of meteorology and dispersion climate have also been conducted (Kelly 1976 and 1977).

5.2.3 Other California KGRA Sites

At least five other KGRA areas in California have been active during the past two or three years and suggest that future development may proceed if

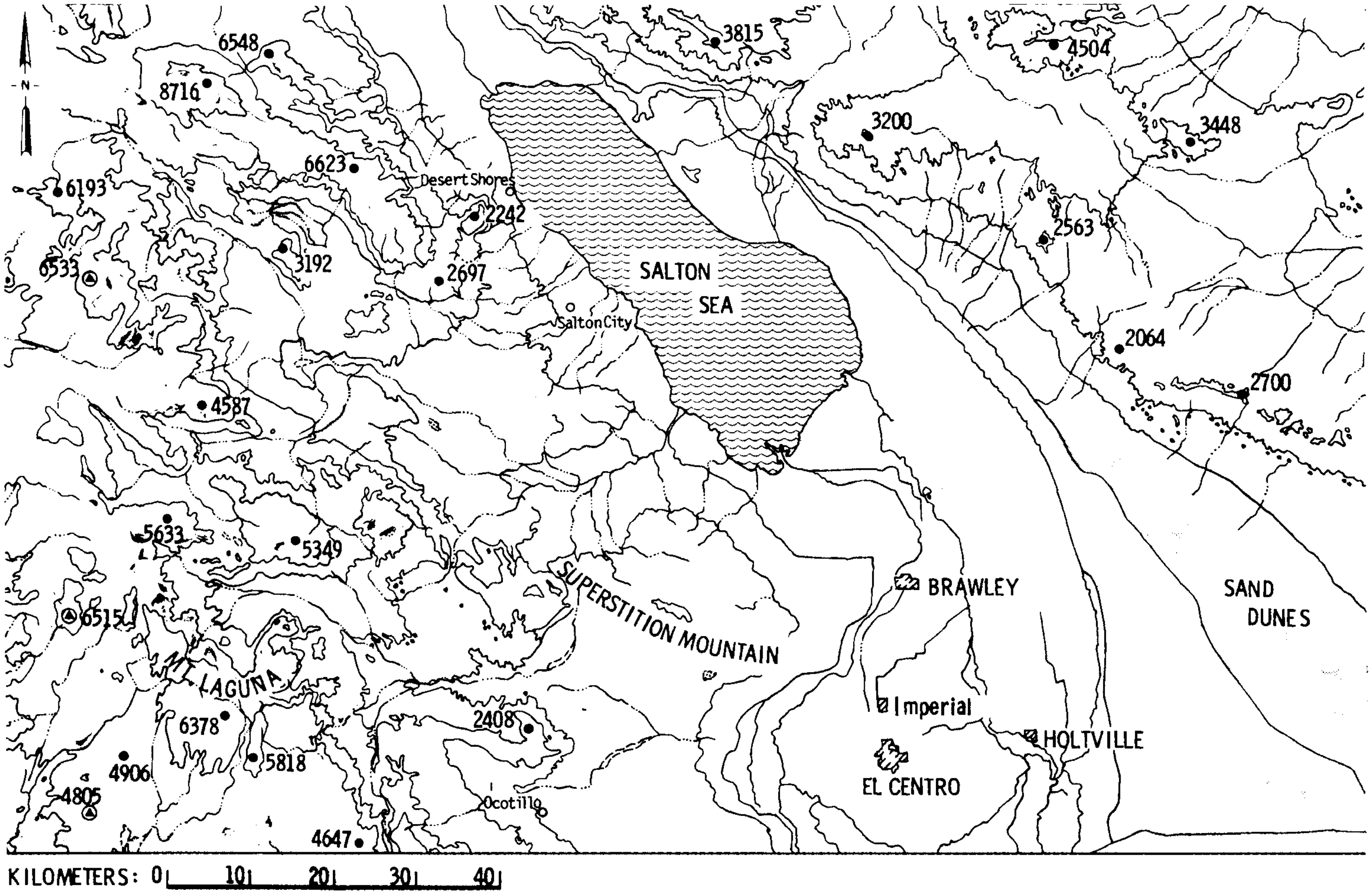


FIGURE 5.14. General Topography of and Surrounding the KGRAs of Imperial Valley, California

various engineering problems are solved. These KGRA sites are being considered in the Geothermal Overview Project for environmental assessment (Phelps et al. 1979).

5.2.3.1 Cosco Hot Spring KGRA

The Cosco Hot Spring site is under jurisdiction of the China Naval Weapons center and is located in the north end of Indian Wells Valley about 50 km north-northwest of the town of China Lake. The valley is located about 25 km east of the Sierra Nevada crest and is situated in the Cosco mountain range (Figure 5.15). Preliminary drilling and testing indicates that this resource is of considerable magnitude, possibly 4000 MW. A 10-MW plant is planned by 1981 (Morse 1979; Smith, Isselhardt, and Matlick 1978; and Leibowitz 1978). The dispersion climate will definitely be influenced by the rough terrain. Local slope-valley winds will be present and during the winter episodes of strong lee waves will occur.

5.2.3.2 Mono-Long Valley KGRA

This KGRA area is located about 40 to 80 km northwest of the town of Bishop, just east of the Sierra Nevada mountains (Figure 5.16). Long Valley is nearly totally surrounded by higher terrain to the west (Sierra Nevada) and to the east and north (Glass Mountain). The dispersion climate of this area, like Cosco Hot Springs KGRA, will be dominated by the surrounding terrain. A study on the climatology of air quality in the Long Valley Geothermal Resource Area was completed in 1977 (Peterson and Palmer 1977). A 60 MW_t (thermal megawatt) plant is planned by 1983 (Morse 1979). Leibowitz (1978) identifies the Mono-Long Valley KGRA as having the largest (~6000 MWe) potential geothermal source in California.

5.2.3.3 Lake City-Surprise Valley

The Lake City-Surprise Valley KGRA is located approximately 40 km east-northeast of Alturas. The area consists of north-south ridges and a desert valley. Leibowitz (1978) indicates that this area ranks fourth (~2000 MWe) in the potential for geothermal resources in the State of California.

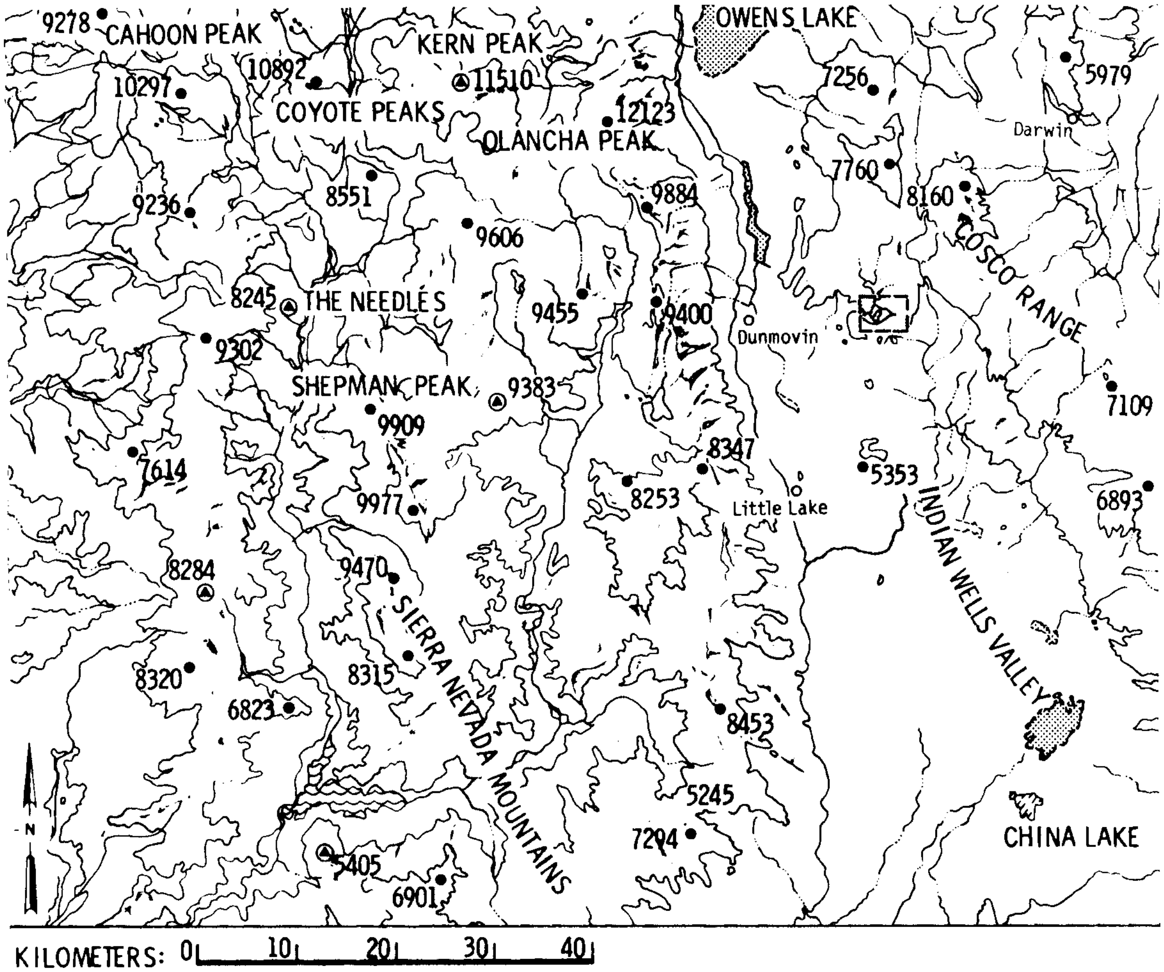


FIGURE 5.15. General Topography of and surrounding the Cosco Hot Spring KGRA, California

5.2.3.4 Wendel-Amedee KGRA

This KGRA is located in northeastern California about 40 km east-southeast of the town of Susanville near Honey Lake. THe area is near a 2300-m mountain and the terrain consists of ridges, valleys, and isolated mountain peaks (Figure 5.17). Dispersion climate will depend upon terrain effects. Stagnation episodes will exist. A 55-MW plant is planned by 1983 (Morse 1979).

5.2.3.5 Lassen KGRA

This KGRA is located just south of Lassen Peak and Lassen Volcanic National Park. Some exploratory drilling and testing is in progress (Smith et al. 1978).

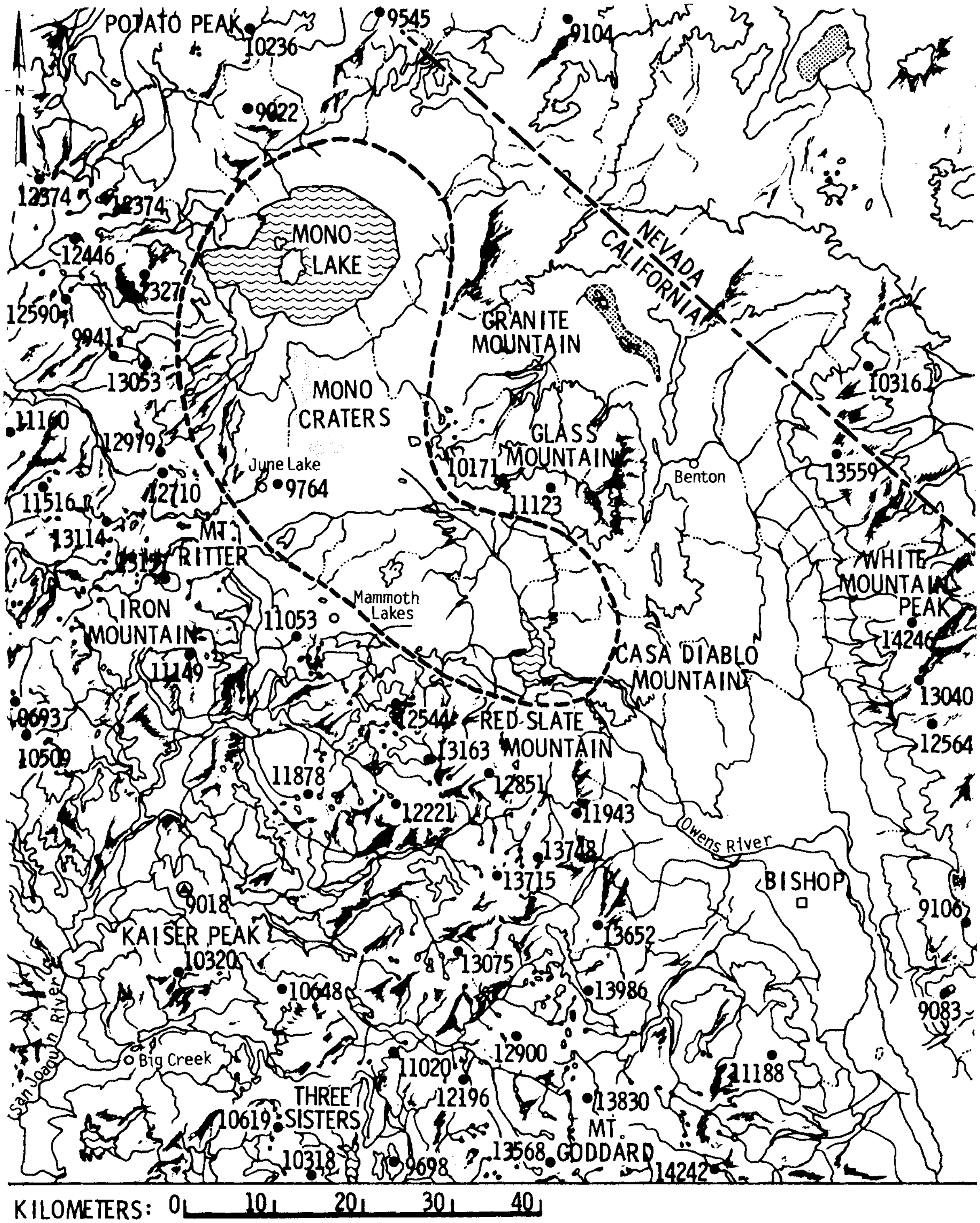


FIGURE 5.16. General Topography of and Surrounding the Mono-Long Valley KGRA, California

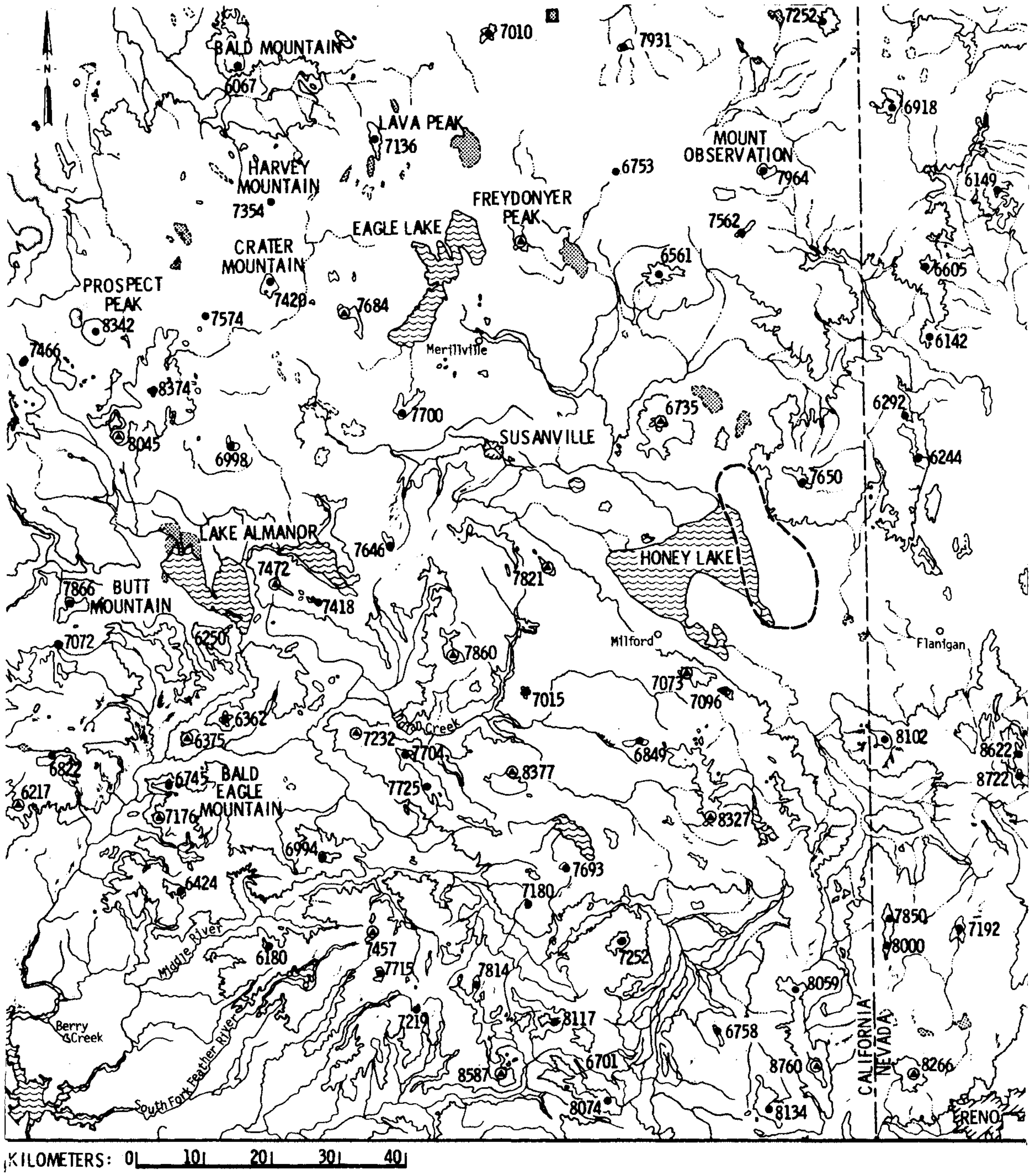


FIGURE 5.17. General Topography of and Surrounding the Wendel-Amedee KGRA, California

5.2.4. Hawaiian Sites

The state of Hawaii was not included in the classification of KGRAs but since the state wants to be energy self-sufficient as soon as possible, Hawaii Electric Light Company (HELCO) is very interested in geothermal energy. Two sites are under investigation: one in the Puna District and the other in the North Kona District.

5.2.4.1 Puna District

The well for a 3-MW demonstration power plant is scheduled to go on line in 1980 and is located on the rift of the east flank of Kilauea Volcano.

5.2.4.2. North Kona District

Exploratory drilling has taken place on the west side of the big island of Hawaii about 10 km east of Kiholo Bay. Dispersion climate for the site will depend on the tradewind regime and the interaction of the synoptic-scale tradewinds and the local land-sea breeze regime.

5.2.5 Idaho KGRA Sites

The Snake River Plain of Idaho is speculated to have more than 25% of the potential geothermal resources of the western United States (Youngquist 1979). However, the evaluation of this highly prospective region is proceeding slowly. The principal area of interest is in the Raft River KGRA where the near-term goal is to produce approximately 15 MW of electricity in a pilot project (U.S. Department of Energy 1978, and EG and G 1976).

5.2.5.1 Raft River KGRA

The Raft River KGRA is located about 40 to 60 km southeast of the town of Burley, Idaho. The site is in the Raft River Valley, a high (~1500 m) desert plain surrounded by higher (2400 to 2800 m) mountain peaks (see Figure 5.18).

Current environmental and meteorological studies have not been available. Generally, the dispersion climate of this area is dominated by strong diurnal variation in temperature, especially during the summer. Frontal or storm activity during the winter result in episodes of strong, gusty winds. Although

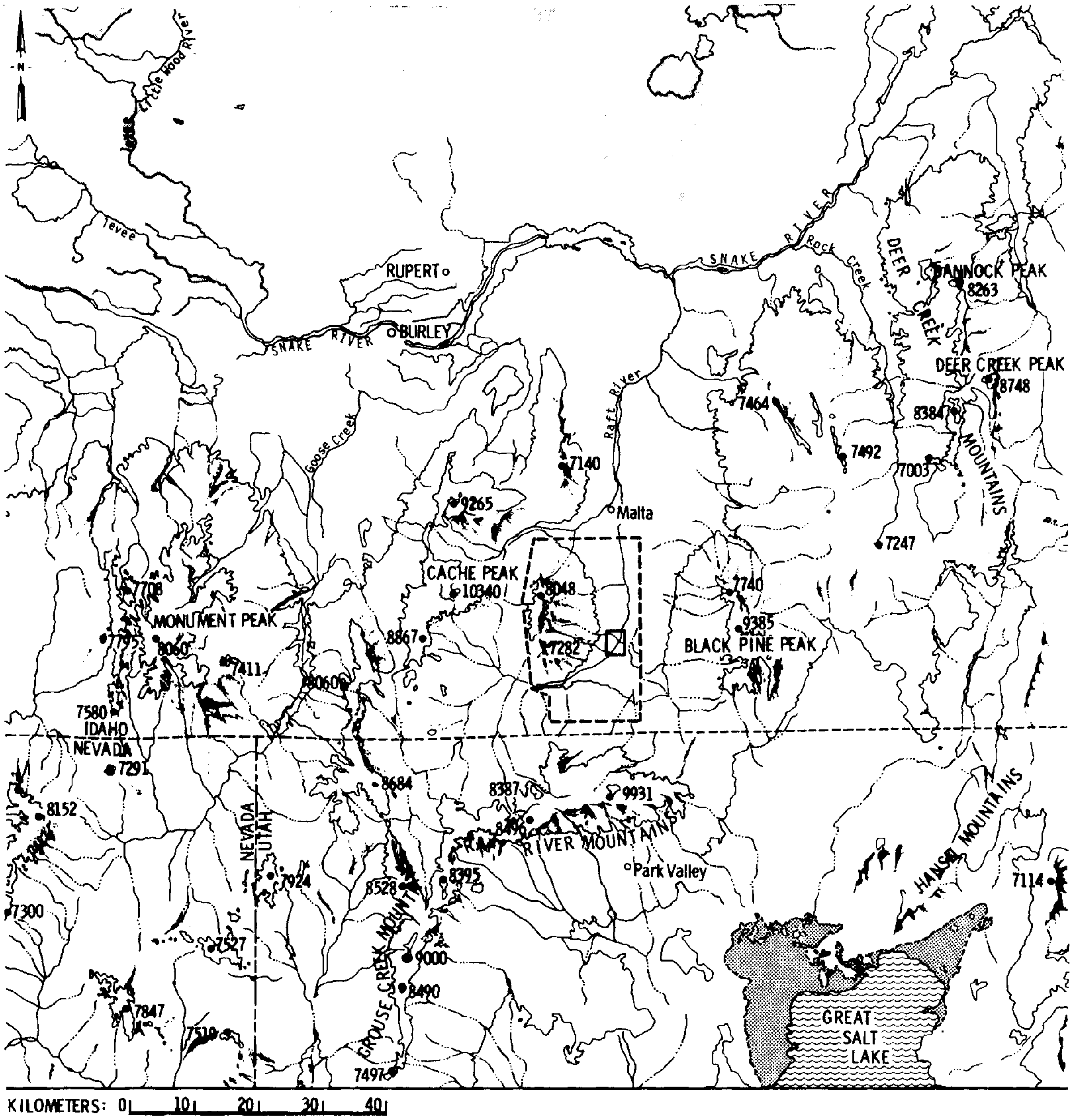


FIGURE 5.18. General Topography of and Surrounding the Raft River KGRA, Idaho

the present development site is in gradually sloping terrain, higher land forms are located about 10 to 25 km away and terrain effects on the local wind regime are expected.

5.2.5.2. Other KGRA Sites

At present, a number of exploratory drillings and work have been conducted at the Crane Creek KGRA and Castle Creek KGRA without significant results. Some activity has also been present near Sugar City, Arco, Preston and Boise (Smith et al. 1978, Youngquist 1979, and Warner 1972).

5.2.6 Nevada KGRA Sites

At least nine KGRAs in Nevada have seen some drilling and development work. These are Dixie Valley, Rye Patch, Gerlach, San Emidio, Wilson Hot Springs, Soda Lake, Beowawe, Monte Neva and Steamboat Springs. At Dixie Valley, the second dry stream field or reservoir in the United States outside of national park boundaries has apparently been confirmed by additional testing (Smith, Matlick, and Asselhardt 1979). Many of the Nevada KGRA sites are located in desert valleys or basins surrounded by north-to-south oriented mountain ridges. Some, like Soda Lake, are on the relatively flat terrain of the Carson Sink.

5.2.7 New Mexico KGRA Sites

The primary objective of the Los Alamos Hot Dry Rock (HDR) Project is to identify regions of hot rock beneath the earth's surface to see if an economical and environmentally acceptable means of extracting a reasonable amount of contained thermal energy can be devised. The DOE-supported HDR geothermal energy project located in the Jemez Mountains (Valles Caldera) approximately 30 km west of Los Alamos is the only large-scale experimental field study of this resource. Initially, this is an attempt to create a pressurized-water circulation loop in hot granite by hydraulic fracturing between two boreholes (Pettitt 1978, Smith 1975, and EOS 1979). In early 1978 the government agreed to match funds with Union Oil and Public Service Company of New Mexico to construct a 50-MW flashed steam electrical demonstration plant for Valles Caldera field site (Smith et al. 1979).

The Valles Caldera is the collapsed center of a large (width ~20 km) inactive volcano. The western flank of this volcano, now known as the Jemez Plateau, extends into the north-south-oriented Nacimiento-San Pedro Mountain range. Thus, the Jemez Plateau is bounded on the east, north and west by mountains that are >3000-m high. The generalized topography surrounding the Valles Caldera is shown in Figure 5.19.

The Plateau surrounding Los Alamos HDR facility ranges in elevation from 2500 to 2800 m. The plateau drains to the south into the Rio Grande. The two major drainages, the Rio del los Vacas and the Jemez River, have eroded major canyon systems into the surface of the plateau. Other numerous small drainages that dissect the plateau account for its rough topography (Figure 5.20). Much of the area is covered with mixed conifer vegetation.

The collection of baseline data for air quality and other environmental evaluations is in progress (Rea 1977). The climate of the area is characterized by localized convective shower activity during the summer and the

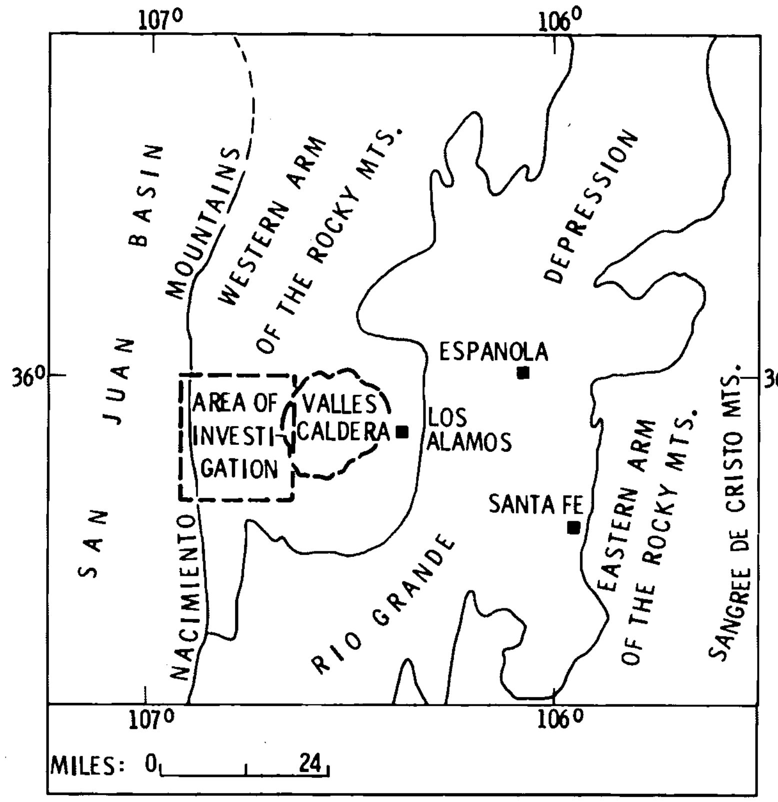


FIGURE 5.19. Major Topographic Features and Area of KGRA within the Valles Caldera (Baca), New Mexico

larger-scale major storms during the winter. Total precipitation (~45 to ~50 cm) generally increases with elevation; however, the irregular terrain results in widespread inhomogeneity in the rainfall patterns. Temperatures of the region are generally mild, with few days exceeding 32°C in the summer and few nights dropping below -18°C in the winter. Nevertheless extreme (>20°C) diurnal fluctuation in temperatures can occur.

The airflow patterns have a distinct seasonal shift. During January the air flow is primarily from the northwest and during August from the southeast. The transition between winter and summer airflow patterns occurs around March. March is also the windiest month of the year. The summer months are relatively calm, indicating that the southeast airflow associated with summer convective activity are not as strong as the northwest flows during the winter.

At present little geothermal development is occuring at the other KGRA sites in New Mexico. However, other HDR sites under regional evaluation are in the eastern states, northcentral New York state, eastern coastal plains and in the states of Washington, Oregon, Arizona[a] and New Mexico.

5.2.8. Oregon KGRA Sites

Geothermal exploratory work was conducted in the Mt. Hood KGRA but primarily for direct heat application. Other exploratory work occurred at the Newberry Caldera, Bully Creek, Crump Geyser, Glass Butte and in the Alvord Desert (Bowen 1972 and Smith et al. 1979).

5.2.9. Utah KGRA Sites

The Roosevelt KGRA has been reported to be one of the hottest geothermal resource areas yet discovered in the United States, and other than its distance from markets, it could be one of the more economically attractive geothermal fields in the country (Rex 1978). Phillips Oil, Utah Power and Light, and Rogers International plan to have a 50-MW power plant in operation in mid-1982. Other private companies plan to have another 55-MW plant in operation by 1982 also (Smith et al. 1978).

(a) See West and Laughlin (1979).

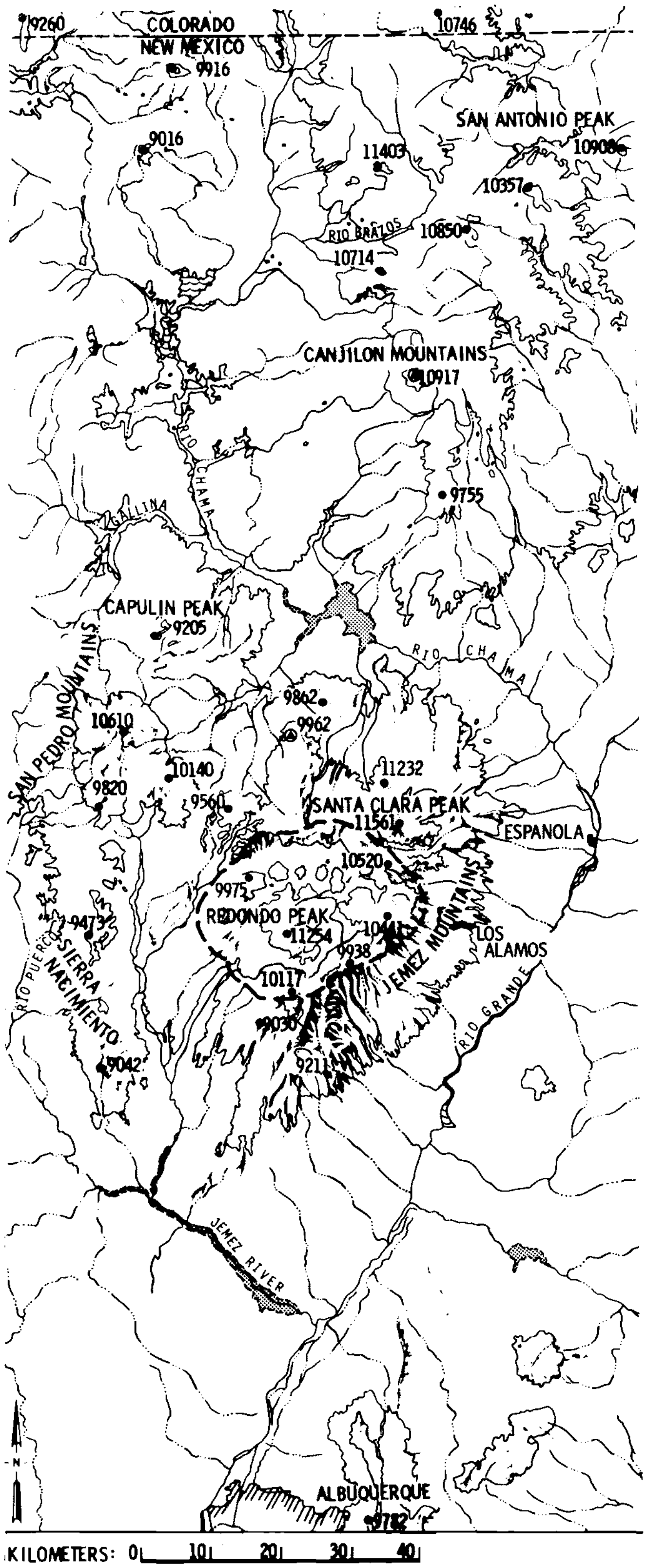


FIGURE 5.20. General Topography of the Valles Caldera (Baca) KGRA, New Mexico

Exploratory work has been accomplished at three other KGRA sites in the same general area. These are Cove Fort, Thermo Hot Springs, and Lund KGRAs (see Figure 5.21).

5.2.10 Eastern United States Sites

Interest has been increasing in the moderate-to-low temperature hydrothermal systems that are likely to exist in many of the sedimentary basins in the eastern United States. Data from the American Association of Petroleum Geologists Survey indicate that there are large areas in the east, associated with major demographic centers, that have useful (49°C to 104°C) temperatures at exploitable depths (900 m to 3000 m).

Conduction-dominated hydrothermal systems in the eastern United States are being evaluated in the Atlantic and Gulf Coastal Plains and in several of the interior basins (Illinois, Michigan, and Appalachians). The conduction-dominated systems are divided into geopressured aquifer systems, such as occur along the Gulf Coast, and simple aquifer systems that occur in most other areas. Convection-dominated systems are being evaluated at Warm Springs, Virginia, and Hot Spring, Arkansas. Such systems may possibly occur in the Champlain Valley and many of the Appalachian-Triassic Basins (Tillman 1979, and Costain, Glover III, and Sinh 1980).

5.2.11. Summary

Use of geothermal resources in the United States is on the threshold of undergoing rapid expansion. The Division of Geothermal Energy (DOE) has currently identified approximately 40 KGRAs as having high possibilities for commercial development. These include the KGRAs that have been listed in Table 5.6 and in the future will involve KGRAs in Hawaii, Nevada, Oregon, and other states.

As indicated in Appendix II, the majority of KGRAs are located in a variety of complex terrain. Table 5.7 indicates the higher priority KGRAs in relation to potential electrical power development. With the possible exception of Imperial Valley, the terrain of the other KGRAs should be an important factor in the location of geothermal facilities and their potential environmental impact.

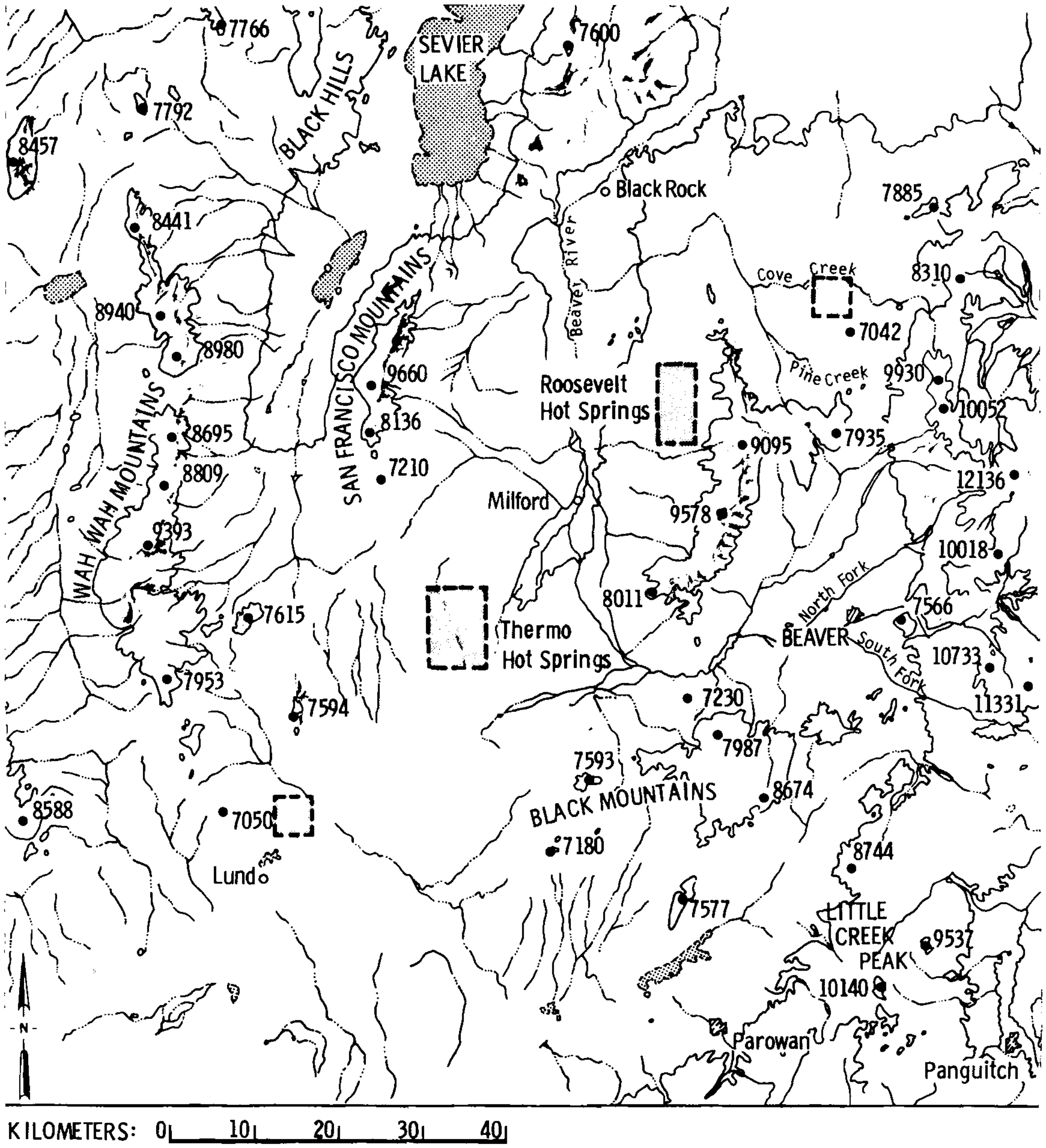


FIGURE 5.21. General Topography Surrounding the Roosevelt Hot Springs, Cove Fort, Thermo Hot Springs, and Lund KGRAs.

TABLE 5.7. Projection of Installed Geothermal Power Capacity Through 1983 (in Megawatts)

State	KGRA	Proposed Power	Terrain
California	Geysers-Calistoga	1248	Ridges, Valleys and Canyons
	Imperial Valley (Six KGRAs)	138	Broad Desert Valley
	Cosco Hot Springs	10	Leeside-Mountain Slope
	Mono-Long Valley	60 (thermal megawatts)	Leeside-Basin or Valleys
	Wendel-Amedee	55	Basin
Hawaii	Puna District (KGRA)	3	Island
Idaho	Raft River	15	Desert Valley (Mountains)
New Mexico	Baca (Valles Caldera)	50	Large Caldera (Mountain)
Utah	Roosevelt Hot Springs	105	Wide Desert Valley

5.3. COAL GASIFICATION AND LIQUEFACTION RESOURCE SITES

Coal still represents 95% of the United States fossil-fuel reserves; yet it supplies less than 25% of the energy now used by the nation. Oil, natural gas, and coal collectively supply about 94% of the nation's energy. Nearly 40% of the oil that the country consumes is imported, and the economic and political hazards of increasing dependency on foreign-originated energy are quite obvious (Hale 1976).

The gas industry and energy planners in both the public and private sector agree that the manufacture of high BTU gas from coal must be rapidly developed to offset declining supplies of natural gas. Substitute natural gas (SNG) from coal is the only long-term solution of the gas shortage because:

1) coal is the nation's most abundant fossil fuel,
2) petroleum fuel stocks are far more costly than coal, and
3) petroleum and natural gas supplies are limited (Seay 1978, Hale 1976).

The distribution of coal resources in the country excluding Alaska, are found in six provinces: Eastern, Interior, Gulf, Northern Great Plains, Rocky Mountains and Coast (Figure 5.22).[a] Four major provinces--Rocky Mountain Northern Great Plains, Interior, and Eastern--contain more than 90 of all coal reserves in the contiguous 48 states. The Northern Great Plains and Rocky Mountain Provinces contain approximately 70% of the coal resources in the four major provinces and most of the nation's low-sulfur coal. Obviously, a significant amount of the coal resources are in or near areas of complex terrain.

Currently there are approximately 40 or more existing or proposed coal gasification or liquefaction projects in the country (Appendix III). These include in-situ gasification, highand low-BTU gasification and liquefaction projects (Figure 5.23). However, there are no full-scale coal gasification or liquefaction plants operating in the United States today. The projects listed in Appendix III represent the major known ventures. A substantial number of other gasification or liquefaction projects may be underway but are not reported here because of their early stage of development or the fact that project details have not been published. This point should be kept in mind when considering the material in the following sections.

A brief examination of the coal provinces, regions[b] and proposed gasification and liquefaction projects will be discussed in the following sections. The Gulf Province will not be included because the majority of the area is generally flat or rolling terrain.

5.3.1. Eastern Province

The Eastern Province is comprised of three regions: Applachian, Pennsylvania Anthracite, and Atlantic Coast (Figure 5.24). The Applachian region has around nine proposed or existing gasification plants. Many of these are pilot plants.

(a) A province, the largest unit used by USGS to define the areal extent of coal resources, is made up of regions on the basis of similarity in the physical features of coal fields, coal quality, and contiguity.

(b) Regions are made up of fields made up of districts. A field is a recognizable single coal-bearing territory; a district is an identifiable center of coal mining operations.

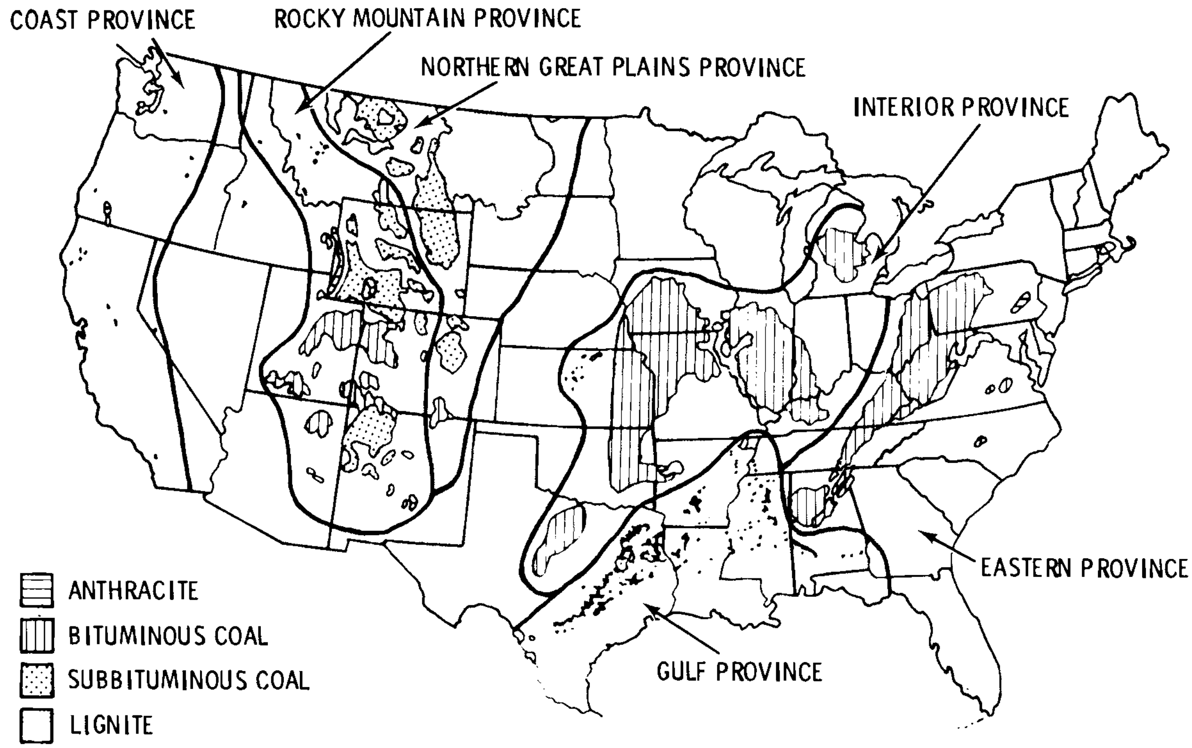


FIGURE 5.22. Distribution of Coal Resources in the United States Excluding Alaska and Hawaii (adapted from Bureau of Land Management 1974:I-47)

The majority of the Appalachian region is mountainous and the general dispersion climate can be poor due to stagnation conditions and the terrain. Obviously the selection of future full-scale gasification plant sites in this region will have to be made carefully especially through Pennsylvania, West Virginia, and Tennessee.

5.3.2. Interior Province

Four regions comprise the Interior Province: Northern, Eastern, Western, and Southwestern (Figure 5.25). A number of gasification plants have been proposed for the Illinois Eastern Region. At least three have been identified for Kentucky.

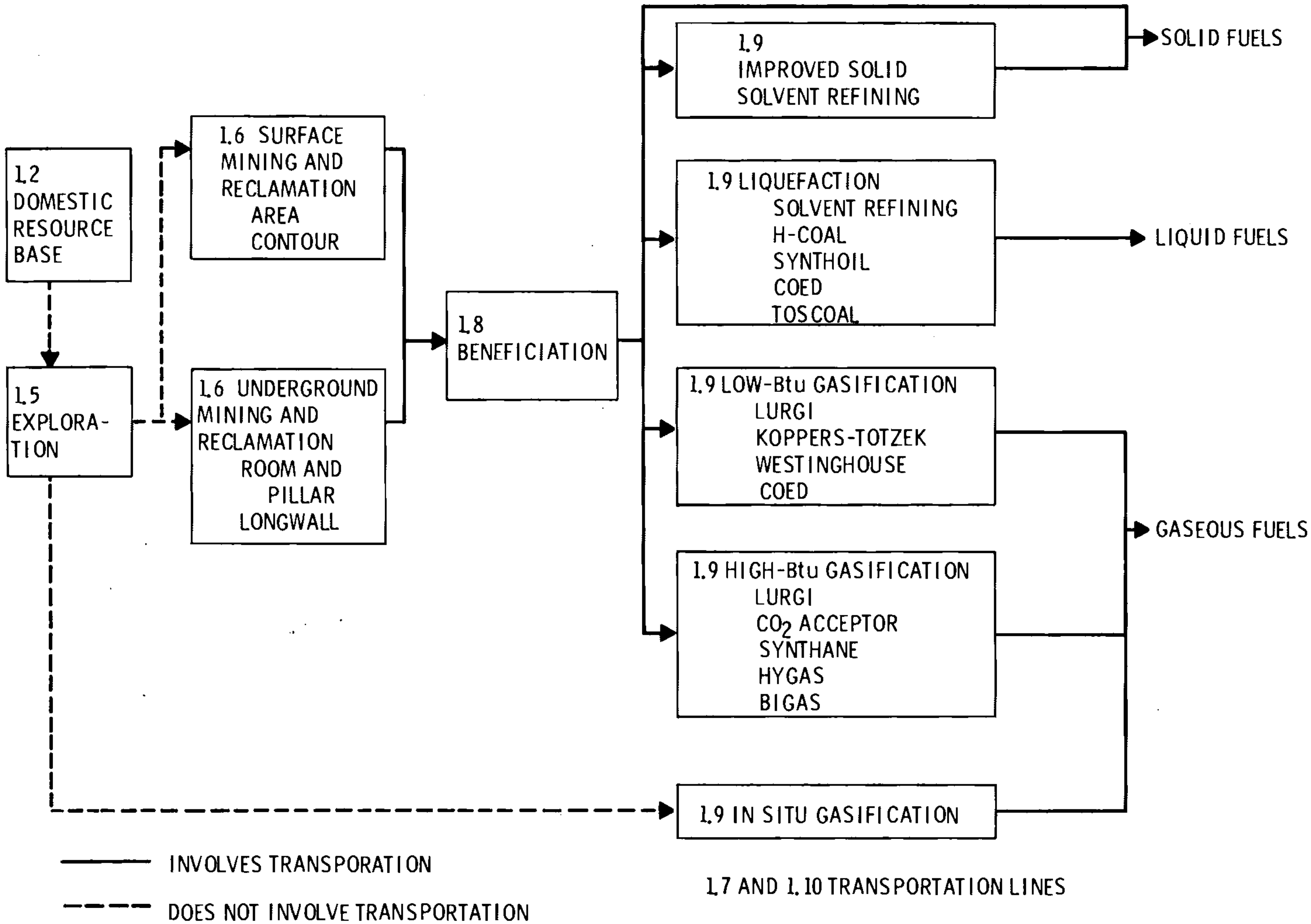


FIGURE 5.23. General Development Plan for Coal Resources

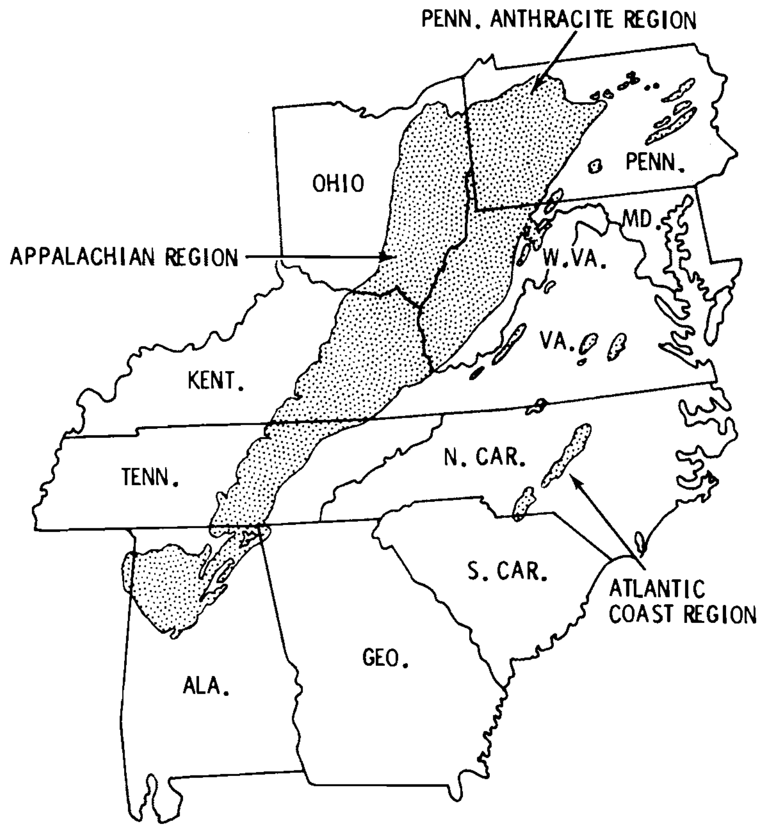


FIGURE 5.24. Distribution of Coal in the Eastern Province (adapted from Bureau of Land Management, 1974:I-47)

5.3.3. Northern Great Plains Province

As illustrated in Figure 5.26, the Northern Great Plains Province, which contains 45% of the remaining coal resources in the United States, is made up of six regions. The two largest regions, Fort Union and Powder River, contain about 1.5 trillion tons of coal, most of which is owned by the federal government. As indicated from Appendix III several gasification and liquefaction projects have been proposed for the Fort Union and Power River Regions. Some activity has also been proposed in South Dakota and eastern Colorado.

The Fort Union Region is generally open and rolling. The plains extend throughout, so this area has few well-defined topographic features. The Missouri River flows through the region and damming of the river has formed two large lakes: Fort Peck and Sakakawea. The Buelah-Hazen area just south of Lake Sakakawea is one major site for gasification plants.

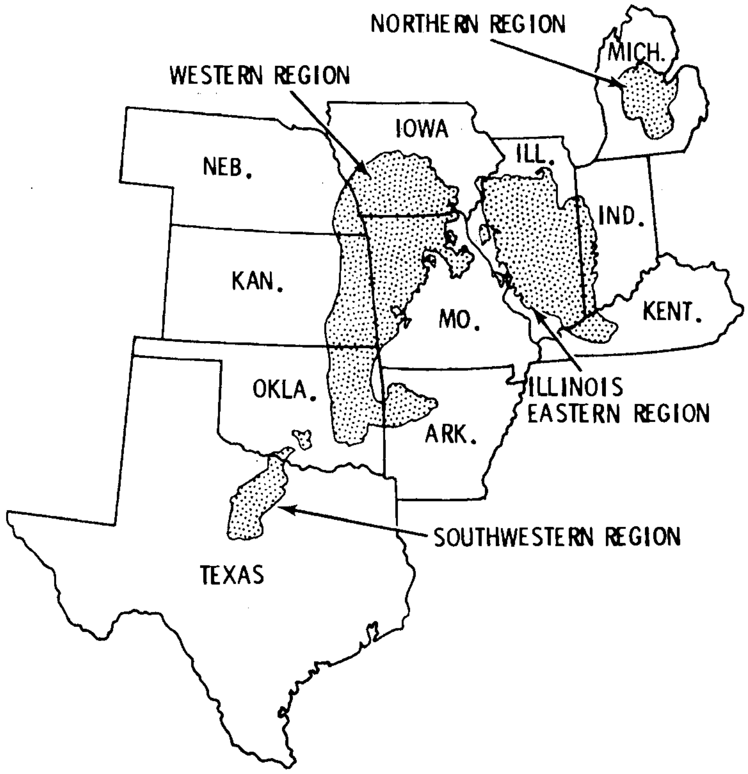


FIGURE 5.25. Distribution of Coal in the Interior Province (adapted from Bureau of Land Management (1974:I-47)

The Powder River Region is a little more rugged than the Fort Union Region but is still open and rolling terrain. In Wyoming, the region is bordered to the east by the Black hills. Two principal rivers, the Yellowstone and Powder, flow through the area.

5.3.4. Rocky Mountain Province

Of the Rocky Mountain Provinces' eight regions the Green River, Uinta, and San Juan River are the largest (Figure 5.27). This province has the greatest variety of topography of any province in the United States. Coals of greatest current interest are subbituminous and low-grade bituminus, found mainly in the southern part of the province and in the Green River and Uinta Regions.

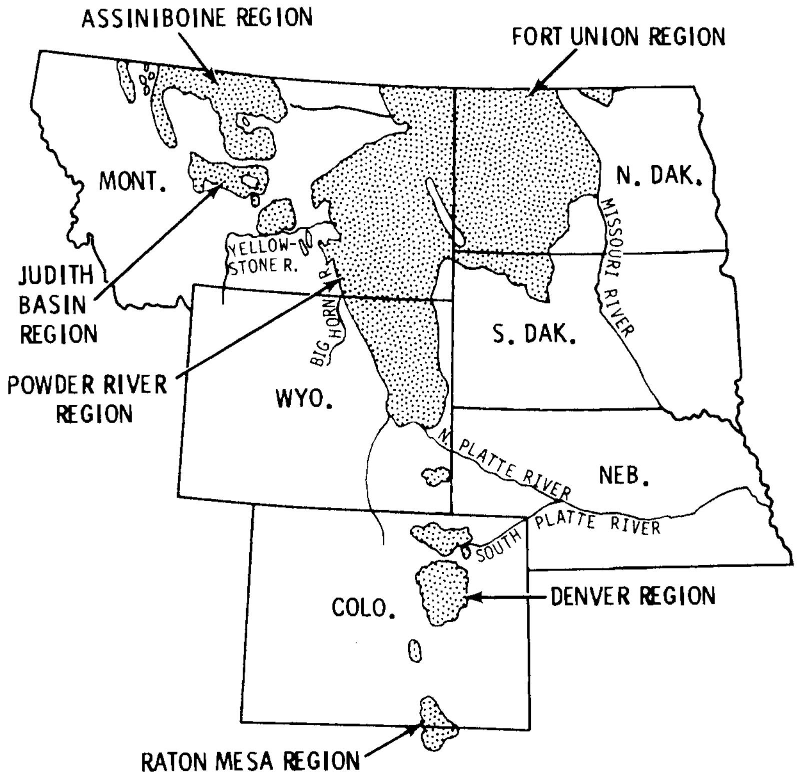


FIGURE 5.26. Distribution of Coal in the Northern Great Plains Province (adapted from Bureau of Land Management 1974:I-47)

A large coal gasification project has been proposed for the San Juan River Region (Smith 1975). If this project is constructed, its location will be on the Navajo Indian Reservation about 40 km south-southwest of Farmington, New Mexico, and the San Juan River. The area is mostly desert with little distinct topographic features. The nearest mountain range, Chuska, is approximately 60 km to the west.

Other gasification projects are planned for north central and northwestern Wyoming (Anon 1979). An in-situ coal gasification project has been operating near Hanna, Wyoming, in the eastern part of the Green River Region since 1972. The topography of the northern part of Wyoming is varied, consisting of relatively large open basins and high north-south mountain ranges.

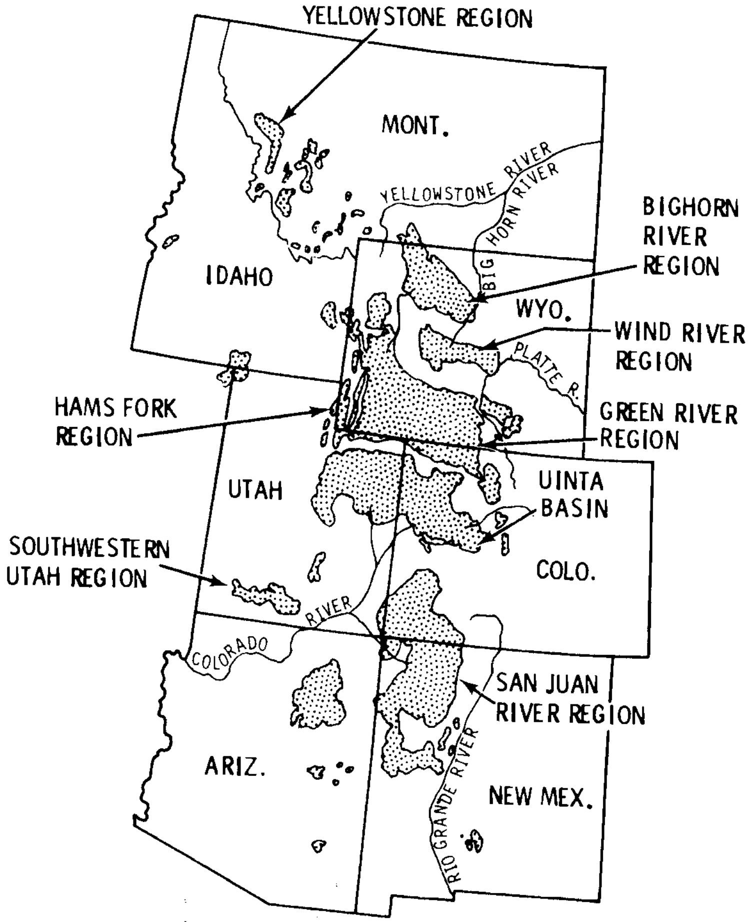


FIGURE 5.27. Distribution of Coal in the Rocky Mountain Province (adapted from Bureau of Land Management 1974:I-47)

5.3.5. Coast Province

A joint coal gasification project, called the Cool Water Coal Gasification Project, is planned for a site near Daggett, California, about 225 km northeast of Los Angeles and 15 km from the desert community of Barstow (Anon 1979). The area is a desert with mountains (1350 to 1900 m) to the north and south about 25 km away.

A pilot coal liquefaction plant is located on the south end of Puget Sound at Fort Lewis, Washington.

5.3.6. Summary

A Federal Power Commission's (FPC) natural gas survey has indicated that the quickest way to increase gas supplies in the next decade will be by coal gasification. According to the FPC report, the coal gasification industry will have from 15 to 63 plants in operation by 1990. Other estimates range from 10 to 30 plants. No commercial plant is expected to be on-line before 1982, with one to two plants per year added thereafter (Hale 1975 and Seay 1978).

The areas of greatest development will be in the Eastern, Interior, Northern Great Plains and Rocky Mountain Coal Provinces. Topography will be an important consideration in siting larger coal gasification and liquefaction plants in many of the regions, particularly, the Appalachian, Powder River, and several of the regions comprising the Rocky Mountain Province. Potential energy facility sites have been investigated for the Rocky Mountain Province by Hinman and Leonard (1977).

5.4. OTHER SITES ASSOCIATED WITH COAL, OIL AND GAS

A large number of other potential candidate sites could be identified with other energy-related development such as, coal-, oil-, gas-fired power plants, refineries, storage areas and ports. In some cases these sites may overlap with the oil shale, geothermal, and coal gasification/liquefaction developments, but it is not in the scope of this report to cover these types of sites in detail.

5.4.1. Electrical Generating Power Plants

The energy policy in the United States has dictated that coal be the primary source of energy generation in the foreseeable future. Many electricity generating plants currently fueled by oil or natural gas may be required to switch to coal.

The distribution of existing coal-and oil-fired power plants is shown in Figure 5.28 (Electrical World 1978-79). Some areas of the country where significant development in coal-fired power plants is taking place are:
a) Montana (Colstrip), b) Southeastern c) Utah (IPP), d) New Mexico (Four Corners), e) Pacific Northwest (Washington/Oregon), and f) the East. The local

and regional air quality changes associated with burning of coal are currently being studied with regard to health effects, changes in concentrations of certain pollutants and changes in visbility. In many of the sites in the West, complex terrain is an important factor in assessing the air quality changes.

5.4.2. Refineries and Fuel Storage Areas

In some cases refineries and fuel storage areas may overlap with oil shale, geothermal and coal gasification/liquefaction developments. The present status of refineries in the United States is presented in Figure 5.29 (O and GJ 1977, IPE 1978, and IPE 1979).

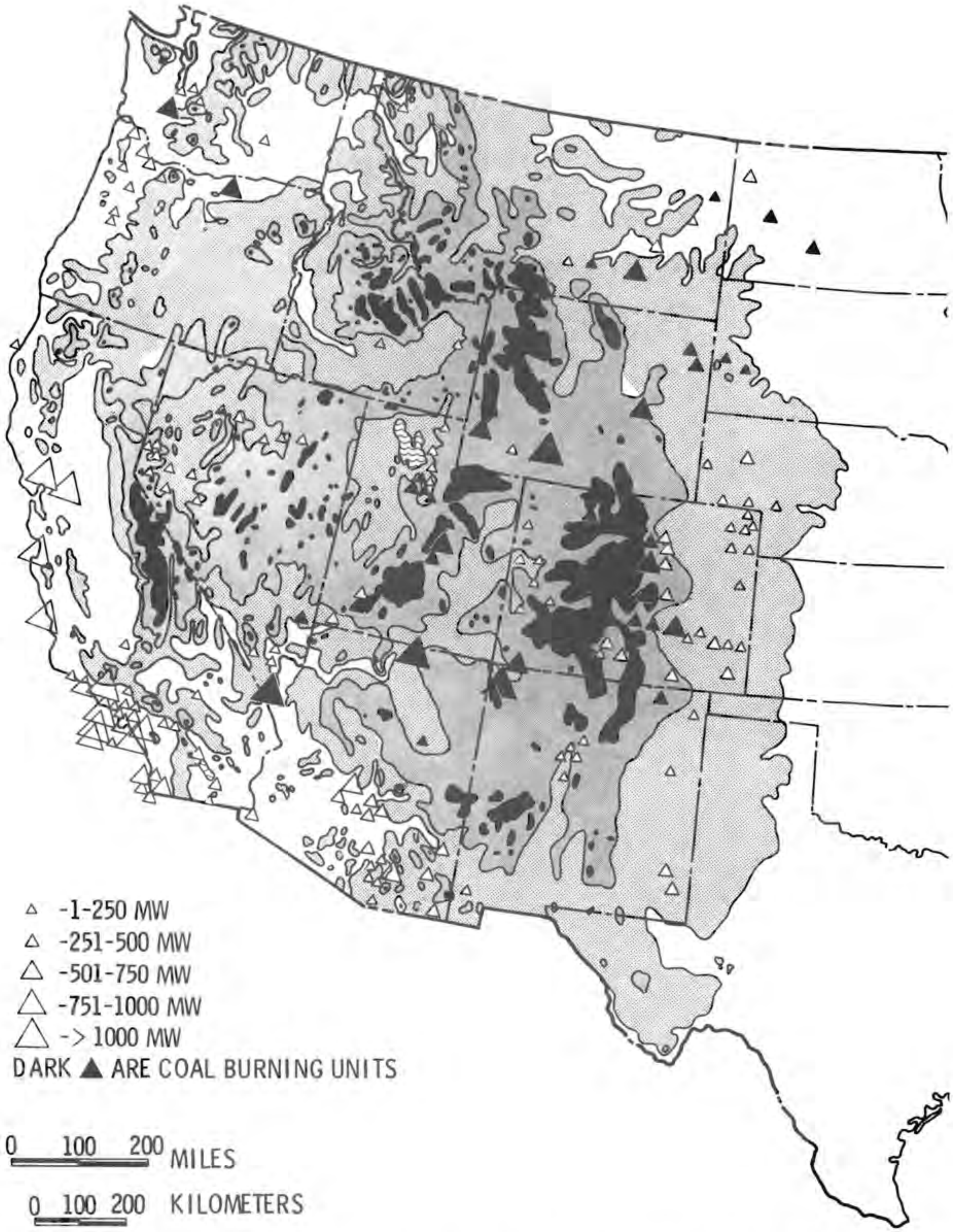


FIGURE 5.28a. Distribution of Existing Coal- Oil- Gas- and Diesel-Fired Power Plants in the Western States

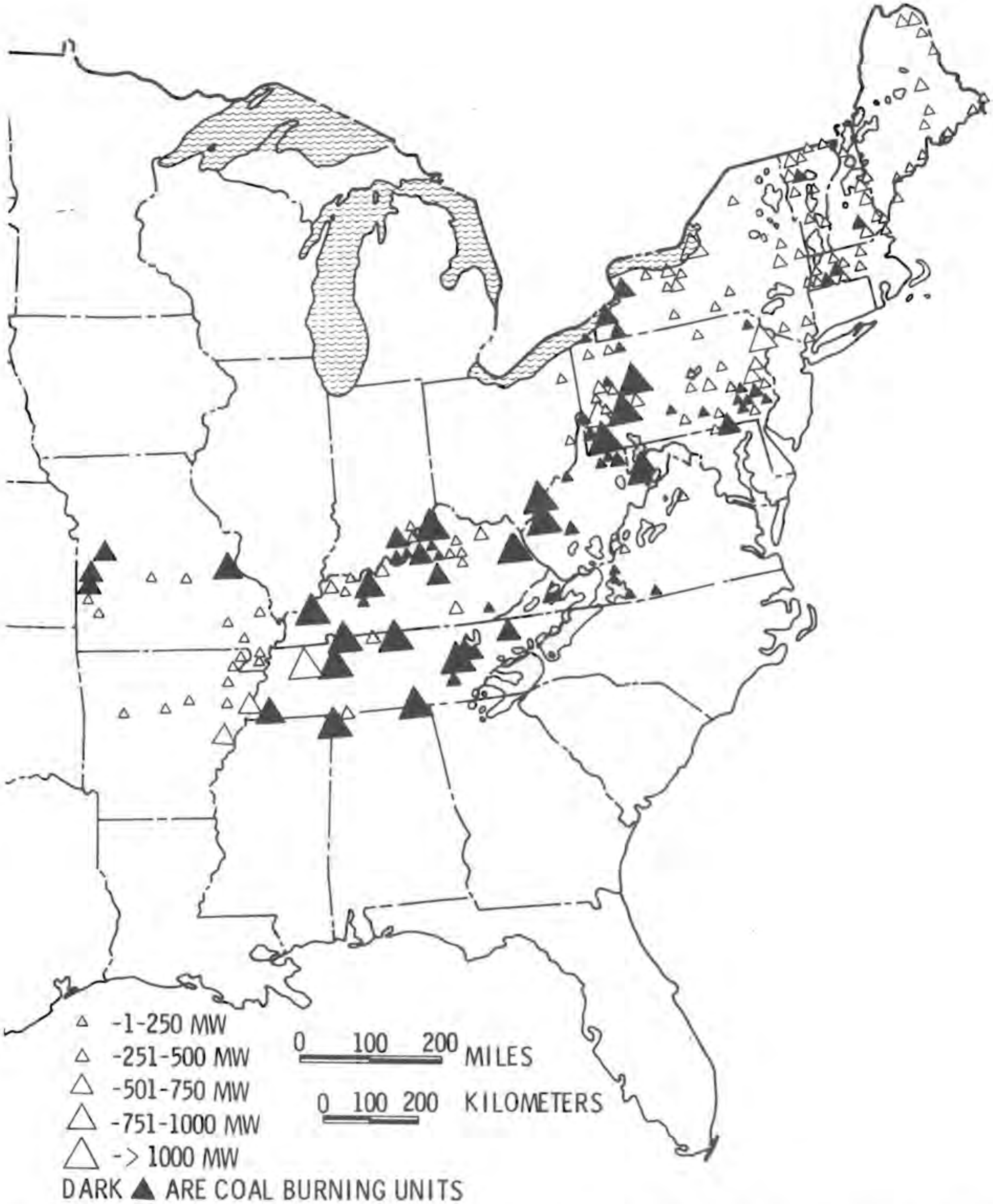


FIGURE 5.28b. Distribution of Existing Coal- Oil-Gas- and Diesel-Fired Power Plants in the Mountainous Eastern States

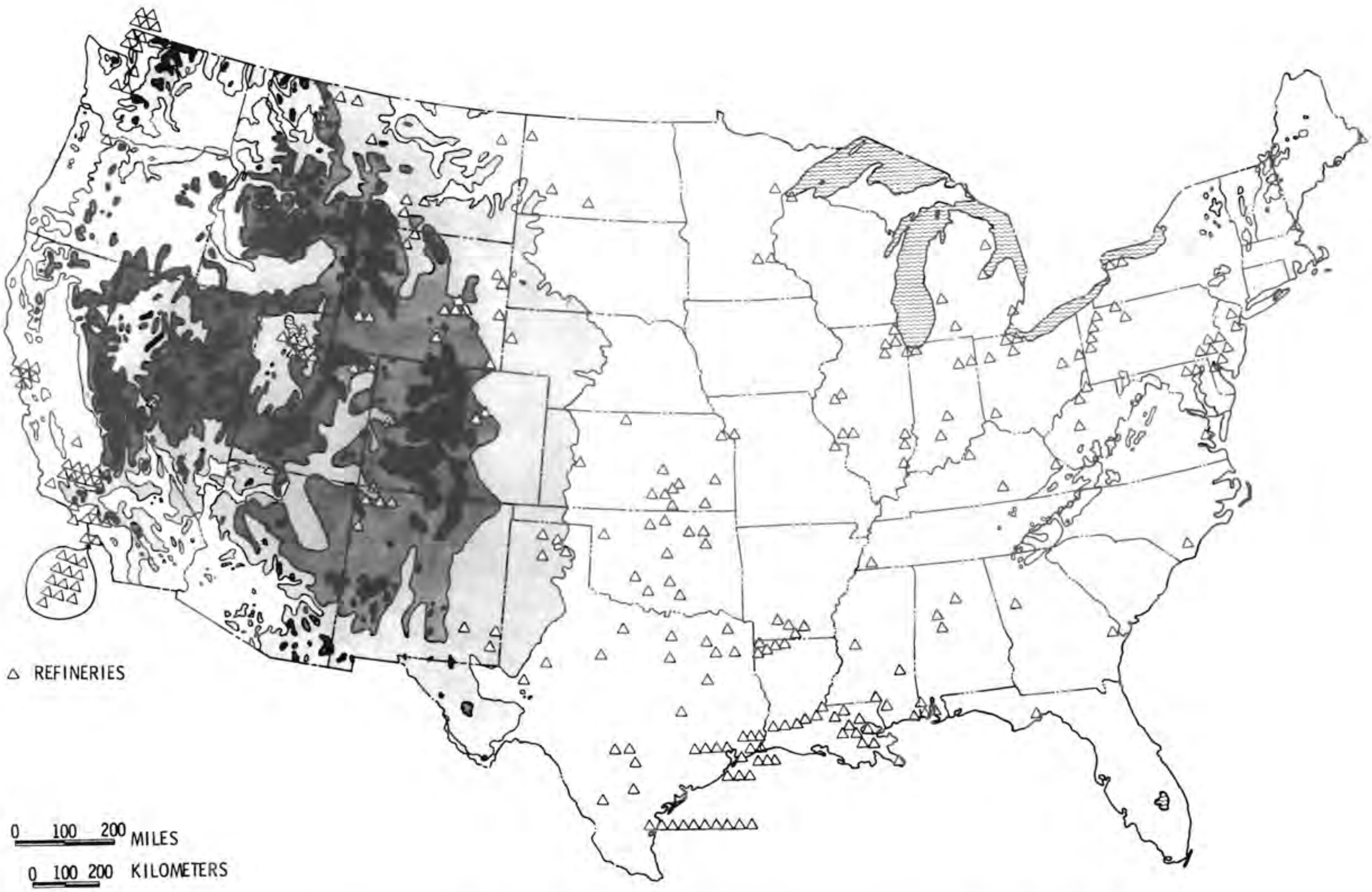


FIGURE 5.29. Distribution of Oil Refineries and Storage Areas in the United States

5.5. OTHER ENERGY RELATED SITES

The last category of sites are open-pit mining, smelters transportation corridors and urban centers in mountainous areas. These sites are involved with energy dissipation as well as energy development and are included since, again, some of these sites may occur in the same general area as the oil shale, geothermal, and coal gasification/liquefaction development.

Figure 5.30 shows the known sites for open-pit mining in the mountainous sections of the United States. These include coal, oil shale, molybdenum, copper, iron, bauxite, tin, uranium, oil shale and zinc. Figure 5.31 indicates the distribution of smelters in the mountainous sections of the United States. These include copper, aluminum, zinc and tin (Bureau of Mines 1975, Aluminum Assoc. 1976, and E and MJ 1980).

Major transportation corridors in mountainous terrain such as interstate highways could provide examples of line-type sources in complex terrain. Cities or moderate-sized urban centers in valleys or basins may be examples of area or volume sources.

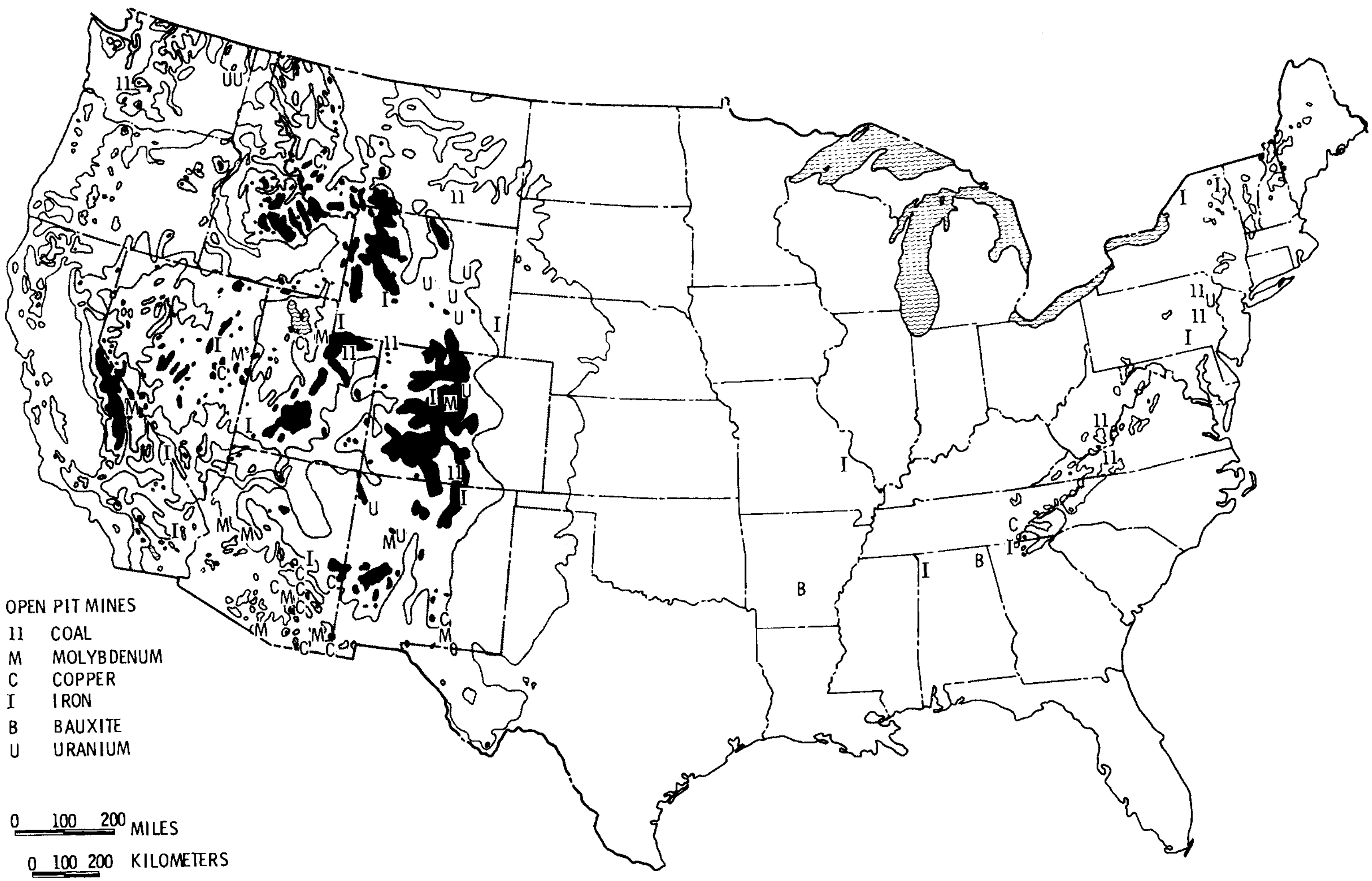


FIGURE 5.30. Distribution of Open-Pit Mining in the Mountainous Areas of the Contiguous United States

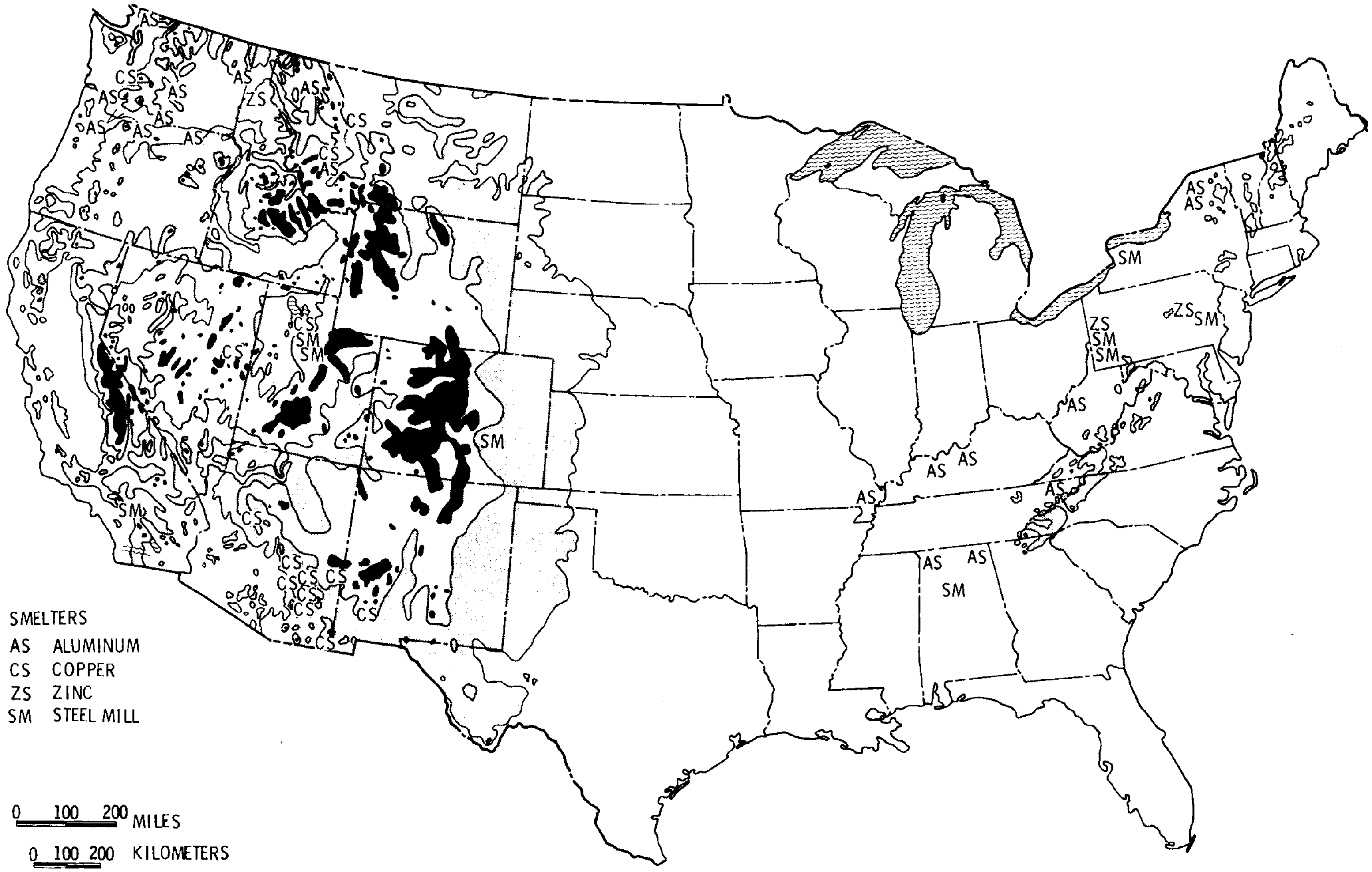


FIGURE 5.31. Distribution of Smelters in the Mountainous Areas of the Contiguous United States

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6.0 DISCUSSION, CONCLUSIONS, AND RECOMMENDATIONS

In the next five years, ASCOT may be directed or elect to extend their research tasks to locations other than the Geysers area depending, of course, on future economic, political, and energy development decisions. To consider some of the possible research options it is important to discuss briefly topics that will be of importance in the selection of new research sites and tasks. After these points are discussed specific research options can be considered.

6.1 SITE SELECTION

A candidate site or sites will certainly depend on whether a particular energy development project is planned to go into operation or undergo expansion. It may also depend on whether this operation poses a possible environmental problem. Thus, candidate sites will most likely be located where potential oil shale, geothermal and coal gasification-liquefaction projects are anticipated (see Section 5.0 for a number of possible sites).

Another approach to site selection would be to site on a generic and theoretical basis. This means that a site would be chosen on the basis of a particular landform (such as a hill, ridge, valley) with the object of investigating in detail and trying to understand as completely as possible the dynamics, kinematics, thermodynamics of the terrain-induced airflow characteristics of that landform.

One of the principal objectives of either approach for selecting a site is to eventually collect data and develop validated models that could be applied to this particular site and possibly to other different landforms.

6.2 TERRAIN UNIQUENESS AND QUANTIFICATION

Terrain uniqueness (i.e., each landform has its own special terrain characteristics) results in interactions with the lower atmosphere producing its own peculiar topo-climate and meteorology. Terrain uniqueness appears obvious when one attempts to catalog the various combinations that the basic components of terrain[a] can assume in the actual world. Even isolated generic landforms or terrain of similar geometry are not very likely to have similarity in all the basic components of terrain.

Terrain uniqueness suggests that qualitative descriptions of particular landforms in a research study are not very adequate for evaluating terrain effects. Terrain quantification or defining the terrain factors that are important to a site appears to have received little attention in many of the studies regarding complex terrain. This study recommends that more emphasis be placed on this facet of the complex terrain problem since it may lead to a better understanding of how topography interacts with the lower atmosphere.

6.3 DEFINITION OF THE RESEARCH PROBLEM

For several prospective sites the basic research problem will probably be defined by the particular site and type of energy development. This means a certain type or combination of meteorological phenomena may have been identified as the principal problem to be investigated. In the case of the Geysers, the nighttime drainage wind phenomenon has been identifed as the initial problem for investigation. This particular local phenomenon may be identified as the main problem in other complex terrain sites but this will depend on the terrain and meteorology of the area.

In some cases the research problem may be well-defined and specific. Once investigated the particular problem is essentially laid to rest. However, a more general approach to the problem would be more scientifically satisfying and complete. In Section 4.0 the types of terrain-induced airflow and diffusion phenomena have been classified under six general headings: 1) dynamic and kinematic effects, 2) local phenomena, 3) boundary layer phenomenon, 4) eddies, 5) turbulence, and 6) plume effects. A suggested general procedure for examining a research problem in complex terrain is shown in Figure 6.1. The basis of Figure 6.1 is shown partially from Figure 4.3 in Section 4.0 and the idea of trying to investigate and understand the relatively less difficult

(a) Basic Components of terrain is defined in Appendix I.

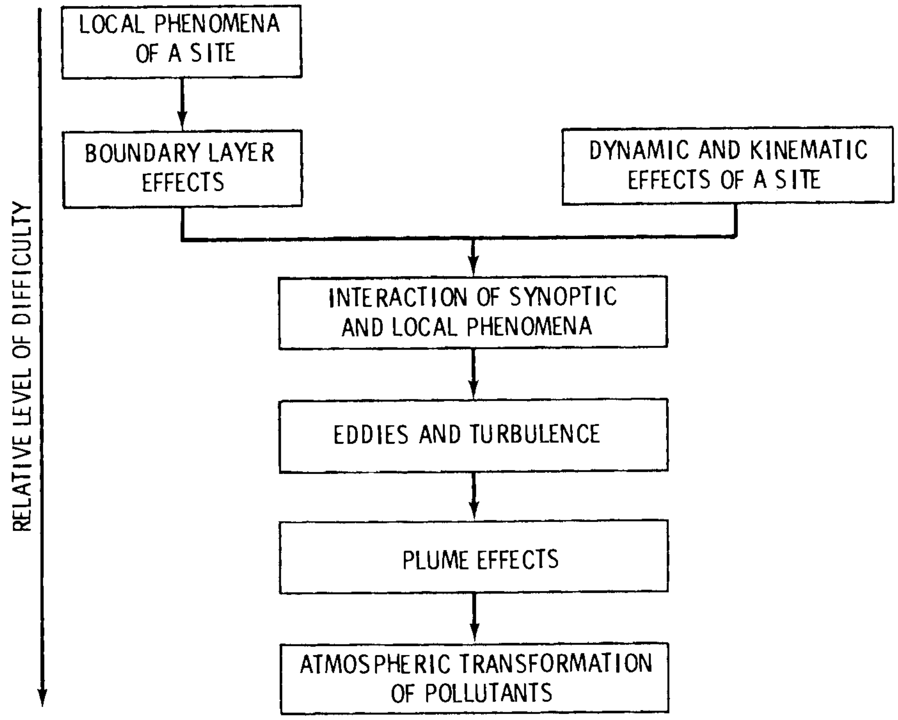

FIGURE 6.1. A Suggested General Procedure for Investigating a Research Problem in Complex Terrain

phenomena (local and boundary layer phenomena) before progressing to the more difficult phenomena such as interactions of synoptic and local phenomena, eddies, turbulence and diffusion. However, this type of research procedure is impossible to do for small-scale research projects where usually one or two aspects of a complex terrain problem are investigated. Yet, in a large-scale project such as ASCOT this research proecedure could possibly be adopted. In fact, this is in some respects how ASCOT has focused on the research problem at The Geysers.

6.4 RESEARCH PLANS

Once a site has been selected for study and a research problem has been proposed, then a specific research plan to study the problem is formulated before proceeding with the work. As indicated by the present information survey the methods of study are essentially: 1) theoretical, 2) numerical, 3) field, and 4) laboratory. A research plan could be designed for a specific site, a selection of sites, or just for model development and application. A general research plan for a specific site will only be considered here. This plan is illustrated in Figure 6.2 and is designed to incorporate the capabilities and advantages of all four research methods.

In Figure 6.2 one of the first phases of a research study at a site is the determination of whether the existing data base is sufficient for conducting initial physical, theoretical and numerical model studies. The initial model studies and field baseline data are to provide the first input data into designing a first effort field program. One approach that has apparently not been attempted is to use the physical model results (and baseline data) for the initial input for a numerical model. Past work suggests that physical models can give an approximate simulation of the velocity fields over various landforms in much greater detail than either field or numerical models (expensive for large complex areas). Results of the initial numerical model can then be utilized for the planning phases of the first field program. As Figure 6.2 indicates, at least two field programs are suggested to provide some minimal seasonal variation to the data. Interaction between subsequent models and data from field programs then take place until the final results are completed.

The ultimate goal of the work in complex terrain is to develop a reasonably "validated" generic model that could be utilized in different landforms for simulating airflow and plume diffusion. Such a goal may be possible in the future in which case the necessity of large field programs as indicated in the dashed box of Figure 6.2 could be eliminated or at least reduced in magnitude. A number of secondary goals will have to be completed before the "validated" generic model is fully developed. Until that time several analytical, physical, and numerical models will be developed and verified that can be applied to different landforms and requirements.

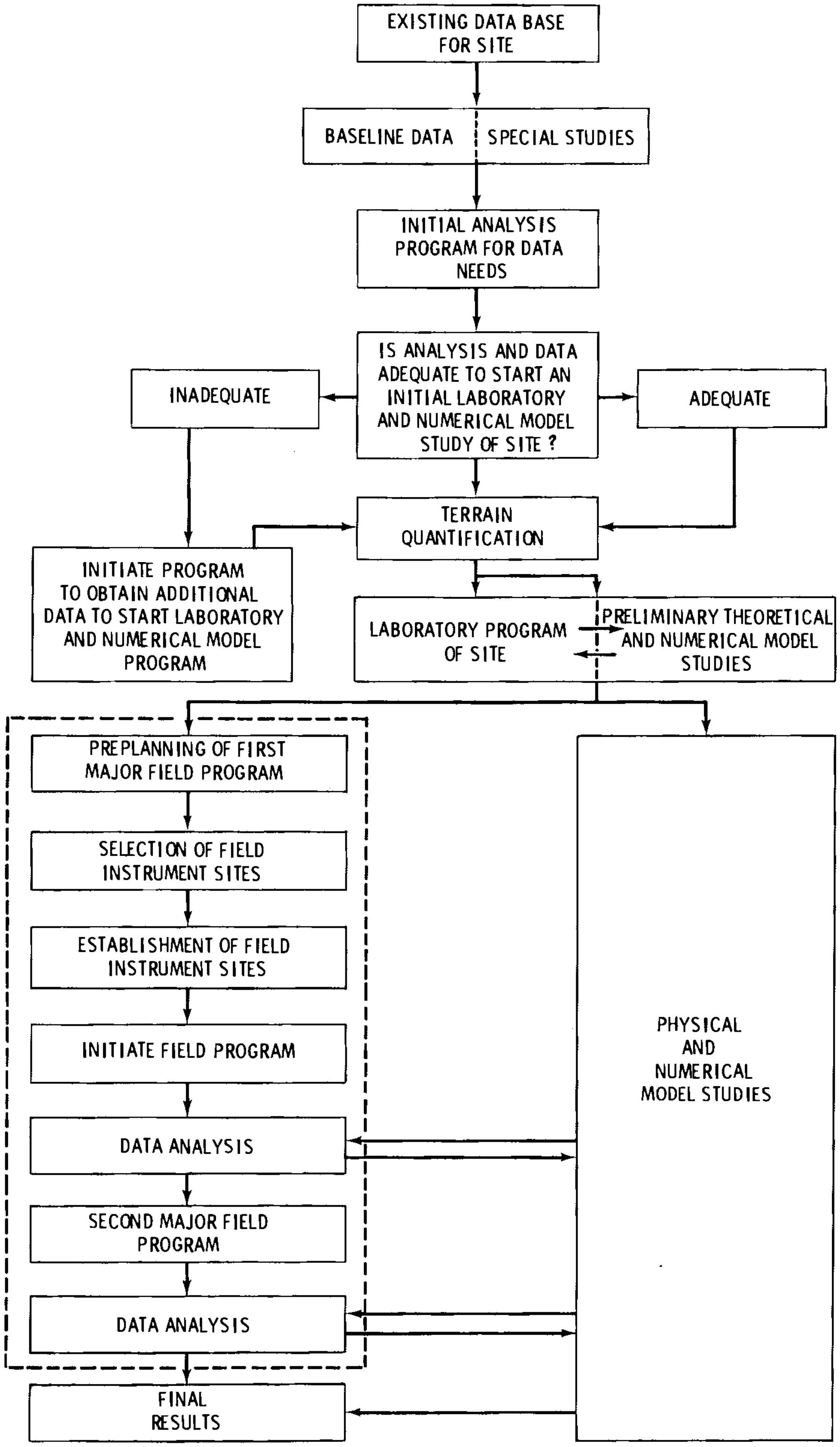

FIGURE 6.2. A General Research Plan for Studying a Site in Complex Terrain

6.5 FUTURE SITES AND RESEARCH OPTIONS

As indicated in Section 5.0, a number of potential candidate sites for the ASCOT program may exist in the future if present projections on fossil energy development are fairly accurate. Political and economic trends will exert their influence on these projected developments. However, the present preliminary evaluation of the various research sites or options will be based primarily on the characteristics of the terrain and their possible contribution to complex terrain meteorological knowledge.

6.5.1 Option 1--Geothermal Resource Sites

At this date ASCOT has completed two phases of field work in The Geysers (Anderson Creek area) and plans are being formulated for one more intensive field effort in the Summer of 1981. Numerical and other modeling efforts are in an initial stage of development.

An important consideration of this present site is whether ASCOT should continue the present studies at The Geysers or consider other complex terrain locations for future work. Some of the advantages of The Geysers site for studies are:

- The logistics of working in the area are not overly complicated by the location and terrain. Also, the availability of the site to a local meteorological group for assisting in the project work is a definite advantage.
- The present site (Anderson Creek area) is a relatively small complex basin-type valley that aids in logistics and collecting data. Yet, the scope of the work and research could be extended to include surrounding topographic features in The Geysers area.
- The location of the project at one site for several years assures a certain amount of continuity in data collection and experience in understanding the meteorology and diffusion processes. In fact, a more general research approach could be assumed as suggested in Figure 6.1.

Several advantages of moving to a different location are:

- Currently and perhaps for several years, geothermal energy development will not constitute a large share of the overall energy development picture; therefore, its impact on the environment will be less than other forms of energy development. It may be advisable to consider a site where the potential environmental problem is in need of immediate consideration.
- In the long-term it is reasonable to collect data at more than one site and to sample a range of different landforms and topo-climates.
- Computer models developed for The Geysers area may not be applicable to other complex terrain areas where the terrain is higher and more rugged. Models need to be developed and tested for a range of landforms and topo-climates to assess their accuracy.
- The present site is not centrally located for all participating laboratories and groups.

In addition to The Geysers area about nine or more geothermal areas could be considered as potential sites as indicated in Table 5.6. Of these sites, the most likely areas of interest would be Imperial Valley, Mono-Long Valley, Lake City-Surprise Valley, Raft River, Valles Caldera, and Roosevelt Hot Springs. In relation to complexity of terrain, the more interesting sites would probably be Mono-Long Valley and the Valles Caldera. Logistics at the Valles Caldera in New Mexico would be relatively less difficult since it is near the Los Alamos Scientific Laboratory.

6.5.2. Option 2--Oil Shale Resource Sites

An obvious second option for ASCOT is the Oil Shale lease sites in Utah and Colorado. At these locations, a three-phase study could be considered if the complexity of the terrain was the principal focus of the project. For example, the U-a and U-b lease sites in northeastern Utah could be investigated first, then the Piceance Basin-Valley Lease sites (C-a and C-b) in Colorado and then some site between the Roan Plateau and Colorado River Valley. This would be an ambitious program but would certainly provide a near ideal data base on the effects of different terrain complexity on meteorology and diffusion. Of course, a less ambitious program would probably focus on one site, say, the Piceance Basin[a] because of the present and future mining and oil shale processing operations. The logistics of these sites will be relatively more difficult than The Geysers because of the size of the area, remoteness, and complexity of terrain.

6.5.3. Option 3--Coal Gasification and Liquefaction Resource Sites

In the near future (5 to 10 yr), coal gasification and liquefaction projects may offer a selection of complex terrain sites (Appendix III). Currently, the possible development in the West appears directed toward the San Juan River Region (New Mexico), Powder River Region (Wyoming and Montana) and Fort Union Region (North Dakota). In general, the current potential sites for development are not located in very complex terrain. However, the status of this option could change quickly if large developments were to occur in the West and the Appalachian region in the East within the next 2 to 5 years.

6.6 CONCLUSIONS AND RECOMMENDATIONS

This last section briefly itemizes some of the conclusions and tentative recommendations that this study has identified for the assistance and guidance of ASCOT in respect to future sites and studies.

1. As the information survey (Sections 2.0 through 5.0) indicates, the data base existing on complex terrain meteorological and diffusion phenomena is fairly extensive. ASCOT should take advantage of this literature and data and use it when possible for guidance and assistance in its data collection, analysis, modeling and field program efforts. However, many

(a) A topic of interest would be to compare the similarities and differences between the Anderson Creek area at The Geysers and the Piceance Basin since they are both basin-like valleys; however, Anderson Creek's exposure is to the east while Piceance Basin's exposure is to the north. Prevailing winds in the lower atmosphere are about the same direction, i.e. southwest and northwest for both sites; yet, the topo-climates are different for the two areas.

data and information gaps still remain in this past data base that provide ASCOT an excellent opportunity to add to and update this complex terrain data base. A wide variety of problems in respect to topometeorology and diffusion in complex terrain still need to be studied for different types of landforms.

2. In the next 5 to 10 years, expansion of oil shale, geothermal and coal gasification-liquefaction development projects in the western and eastern states will provide a range of candidate sites. These sites will provide an opportunity to examine the effects of different landforms on the lower atmosphere.

3. Quantification of the basic components of terrain in problems of topometeorology and complex terrain diffusion have been lacking in past studies. This aspect of the complex terrain problem should be examined more frequently and carefully in future studies.

4. The airflow and diffusion characteristics of different landforms differ as a function of topographic details and meteorology. The concept of studying the airflow patterns for different generic terrain is useful for estimating the general flow characteristics of a particular terrain problem but the details of topometeorology and diffusion will be site-specific for each type of landform because of "terrain uniqueness". Thus, it is suggested that ASCOT consider studying a range of different landforms.

5. An examination of the potential candidate sites for ASCOT suggests at least two options. A third option will depend on how rapid coal gasification-liquefaction and other related technologies develop in complex terrain areas of the country. The first option is to stay at the Geysers area and expand the study program. The second option is to move to a second site. Currently the Oil Shale Resource Area of Utah and Colorado contains some interesting sites since they exhibit a range of complex terrain that, when studied, would add substantially to the complex terrain data base. Other possibilities would be the geothermal resource areas of Mono-Long Valley (California) and the Valles Caldera (New Mexico).

6. A long-term (5 to 10 years) but flexible research plan should be adopted by ASCOT and followed in selecting and conducting field and model studies at a number of sites. Initially this plan should use the advantage of various modeling efforts and field program data. Research should be directed toward understanding local topometeorology and terrain effects while proceeding to the understanding of more complicated aspects of synotpic-local flow interactions, turbulence and plume diffusion and chemical transformations. One of the principal objectives will be to develop a validated operational generic model that could be utilized for different landforms and meteorology. Until the generic model is developed an intermediate objective will be to develop operational models for specific purposes and with specific limitations.

APPENDIX I

CLASSIFICATION AND QUANTIFICATION OF LANDFORMS

The qualitative and quantitative systems for describing and classifying terrain are generally of four types:

- geomorphic classification of singular or isolated landforms
- geomorphic classification of different or similar landforms existing together in a specific area
- quantitative or objective systems for describing and classifying terrain
- computer generated topographic gridded displays.

Geomorphology, or the study of landforms, provides a general classification of terrain types that is generally used when considering the terrain effects of solitary landforms on air motion and transport-dispersion. Some of the landform types that one encounters in the real world are listed and defined in Table I.1 (Finch and Trewartha 1949, and Bradshaw, Abbott, and Gelsthorpe 1978). Thus, an isolated hill is considered a generic landform because its particular geometry affects on-coming air motion in a characteristic mode that is not duplicated by other types of landforms.

Landforms exist on various horizontal and vertical scales; consequently, their effects on atmospheric motion also exist on various horizontal, vertical and time scales (see Section 4.0). Thus, in this respect, each particular landform differs from another even though they may be of the same generic type, e.g., a cliff of small vertical dimensions will not affect the oncoming air as much as a cliff of large vertical dimensions. Obviously, the generic concept of terrain has limitations unless additional quantitative information such as heights, widths, slopes, distances are given to the landform such as illustrated in Figure I-1.

One of the principal utilities of this concept is in recognizing that different generic landforms affect air motion in different characteristic modes and that certain data can be extracted from such studies. For example, laboratory and numerical simulation of airflow over model generic landforms has provided useful information for many site-specific studies.

TABLE I.1. Definition of Some Common Geomorphic or Generic Types of Isolated Landforms

Plateau	- Tabular uplands having a relief of more than 150 m. Some plateaus may terminate in escarpments that make them appear high from at least one side. They may be entrenched with narrow valleys and canyons but the interstream areas are broad and flat topped.
Peneplain	- A plain of low relief with undulating surface and occasional hill remnants, characterized by broadly open valleys.
Cliff (Escarpment)	- A step-like topographic feature that are primarily the result of faulting and water erosion. Generally, their faces are scarred by ravines and sharp promintories. The relief may vary over a considerable range of heights.
Terrace (Benches)	- Series of small escarpments usually of low relief.
Mesa	- A plateau of small-to-moderate size and tabular form (flat top and steep sides). Features of the same origin but smaller sizes are called buttes.
Hill	- Hills are topography that a) are so dissected that much of the land is in considerable degree of slope, b) have uplands of small summit area, and c) have local relief of more than 150 m but less than 600 m.
Isolated Mountain	- Mountains, like hills, are distinguished from plains and plateaus by the smallness of their summit areas and the large proportion of their surface in steep slopes. In general they have more rugged contours and their surface features are more complicated in pattern. The relief is often above 600 meters.
Ridge	- A ridge is an arrangement, usually somewhat linear, of relatively low-lying (<600 m) peaks and their canyons or valleys.
Mountain Ridge (Range)	- Similar to a ridge but in reference to high-lying (>600 m) peaks and their valleys.
Pass (Saddle)	- Passes are the more accessible, and usually the lowest notches across a ridge or mountain ridge.
Gaps	- Narrow notches that have been cut through ridges by major water streams, usually of low or moderate relief.
Gorge	- Relatively long narrow notches that have been cut through mountain ridges by a major river.

TABLE I.1. (continued)

Valley	- The elongated depression caused by water or ice erosion or faulting. Consists of a head (proximal end) that is of higher relief than the mouth (distal end). They may be V-shaped or U-shaped depending on the origin.
Canyon	- Stream-eroded valleys of the narrow and steep-sided form typically found in the semi-arid west. Slopes of canyons may be rather steep.
Basin	- A basin is generally a low-lying plain either partially or completely enclosed by surrounding mountains or ridges. They may vary greatly in area.
Bay	- An irregular formation in a shoreline where a slight or large inundation permits the sea to enter them for some distance. Bays may be associated with beaches, spits, hooks, terraces, escarpments and other features.

Landforms are seldom completely isolated in nature from other types of landforms. Most complex terrain is a conjunction of different or similar landforms existing together in a specific area. Sometimes a general description of such terrain might be adequate. For example, one could state that an area exhibits evenly-spaced, parallel, rolling hills with moderate slopes. Although the description gives basic background information, some tasks could require a more quantitative description. An example of one type of area classification of landforms is shown in Figure I.2. In this case, generic terrain descriptions are provided with additional general information on slope, local relief and profile type (U.S. Geological Survey 1970).

In cases where the geomorphic and generic descriptions of terrain may be still inadequate for a task, then a quantitative description system could be applied. These systems have been principally derived for military applications but may be useful for other tasks. Basically, the technique is to make actual estimates of certain basic components of the actual terrain. These basic components are grouped into five classes:

- Surface geometry
- Composition and consistency of ground
- Vegetation

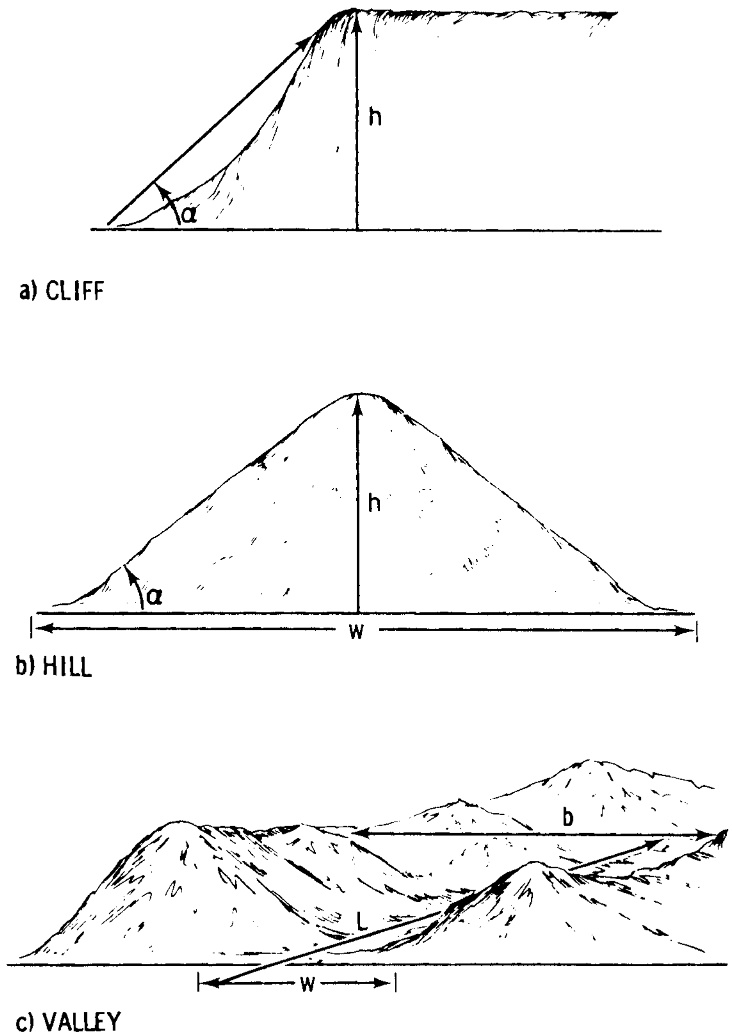

FIGURE I.1. Example of Generic Landforms with Dimensions Specifying Certain Quantitative Aspects of the Terrain: a) cliff: h--height α--slope b) Hill: h--height, α--slope, w--base c) Valley: h--ridge height or depth of valley, w--width at base of valley, L--length of valley, b--width at ridge height.

- Hydrography
- Cultural factors

Techniques for classifying all five components of the terrain are available but this report will only consider the surface geometry (Carr and Lopik 1962).

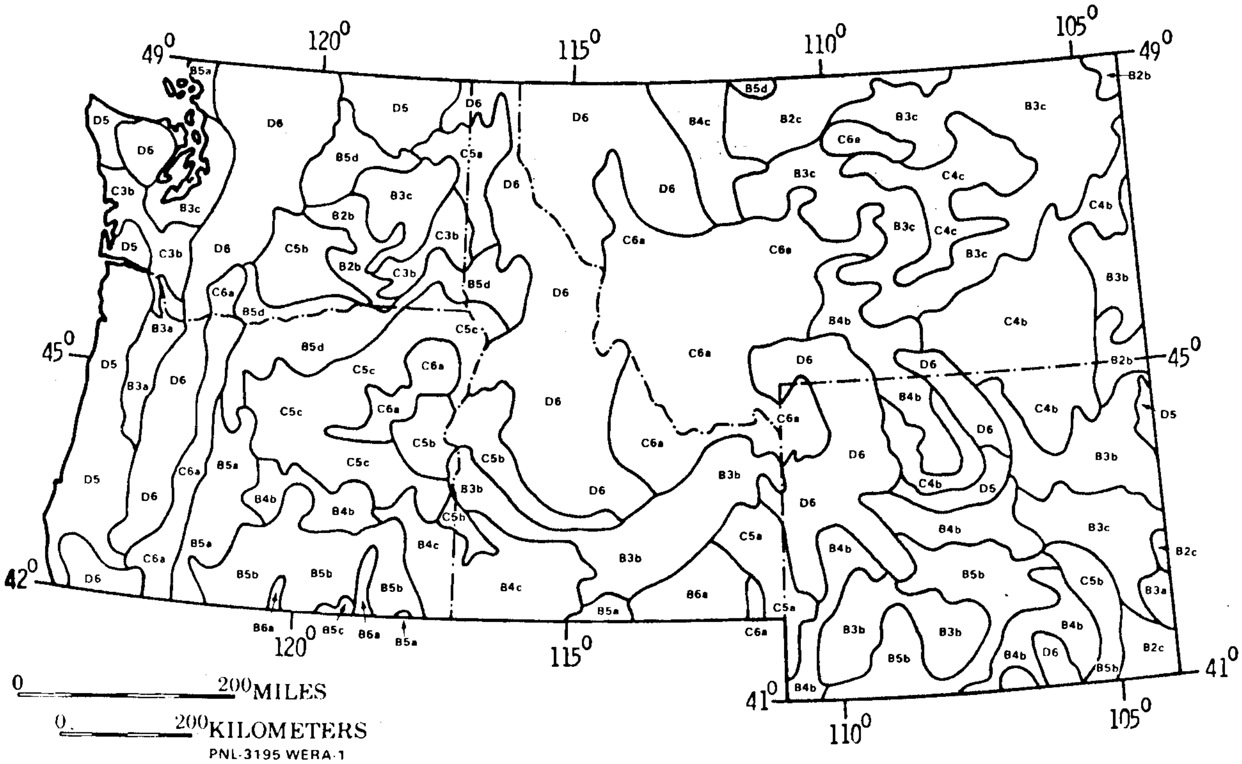

FIGURE I.2a. Classes of Land Surface Forms

Surface geometry concerns the physical form or shapes comprising the earth's surface and their interrelationships. The basic component of surface geometry can be described in terms of the following measurement categories:

- Plan and Profile Measurements--that describe: orientation of topographic highs and/or lows, a real occupancy of topographic highs and/ or lows, planar shape of topographic highs and/or lows, and cross-sectional shape of topographic highs and or/lows.
- Slope Measurements--that describe the inclination or deviation from the horizontal of surfaces forming topographic features or areas.
- Relief Measurements--which describe the vertical dimensions of topographic features or areas.

LEGEND

PLAINS

A1	FLAT PLAINS
A2	SMOOTH PLAINS
B1	IRREGULAR PLAINS, SLIGHT RELIEF
B2	IRREGULAR PLAINS

PLAINS WITH HILLS OR MOUNTAINS

A,B3a,b	PLAINS WITH HILLS
B4,a,b	PLAINS WITH HIGH HILLS
B5a,b	PLAINS WITH LOW MOUNTAINS
B6a,b	PLAINS WITH HIGH MOUNTAINS

OPEN HILLS AND MOUNTAINS

C2	OPEN LOW HILLS
C3	OPEN HILLS
C4	OPEN HIGH HILLS
C5	OPEN LOW MOUNTAINS
C6	OPEN HIGH MOUNTAINS

HILLS AND MOUNTAINS

D3	HILLS
D4	HIGH HILLS
D5	LOW MOUNTAINS
D6	HIGH MOUNTAINS

TABLELANDS

B3c,d	TABLELANDS, MODERATE RELIEF
B4c,d	TABLELANDS, CONSIDERABLE RELIEF
B5c,d	TABLELANDS, HIGH RELIEF
B6c,d	TABLELANDS, VERY HIGH RELIEF

SCHEME OF CLASSIFICATION

SLOPE (1st LETTER)

- A >80% OF AREA GENTLY SLOPING
- B 50-80% OF AREA GENTLY SLOPING
- C 20-50% OF AREA GENTLY SLOPING
- D <20% OF AREA GENTLY SLOPING

LOCAL RELIEF (2nd LETTER)

- 1 0 TO 30m (1 TO 100 ft)
- 2 30 TO 90m (100 TO 300 ft)
- 3 90 TO 150m (300 TO 500 ft)
- 4 150 TO 300m (500 TO 1000 ft)
- 5 300 TO 900m (1000 TO 3000 ft)
- 6 900 TO 1500m (3000 TO 5000 ft)

PROFILE TYPE (3rd LETTER)

- a >75% OF GENTLE SLOPE IS IN LOWLAND
- b 50-75% OF GENTLE SLOPE IS IN LOWLAND
- c 50-75% OF GENTLE SLOPE IS ON UPLAND
- d >75% OF GENTLE SLOPE IS ON UPLAND

FIGURE I.1b. Legend to Figure I.2a

Table I.2 shows some of the surface geometry measurements that could be derived under each of these categories. Details on how the measurements are performed are given in the reference.

One of the surface gemoetry measurements that could be useful in classifying desert terrain is the characteristic plan-profile. The characteristic plan-profile was defined by Van Lopik and Kolb (1958) in work done to describe terrain features in desert areas. The characteristic plan profile (CPP) is a semi-quantitative method that provides a framework to which values of slope, relief, and slope occurrence can be assigned to present an easily conceived mental image of the landscape. Figure I.3 is a graphic representation of the characteristic plan-profile method. A method to quantitatively produce the CPP has been proposed by W. E. Graban (Van Lopik and Kolb 1959).

During the past five years there has been an increase in the use of computers to generate two-or three-dimensional gridded displays of topographic features. These terrain codes have been useful in making estimates on

TABLE I.2. Classification of Selected Surface Geometry Measurements and Properties

Plan and Profile Measurements[a]

Characteristic Plan-Profile	Texture
Statistical Slope Orientation	Roughness
Grain	Valley Width
Slope Direction Change	Elevation - Relief Ratio
Valley Spacing	Hypsometric Integral

Slope Measurements[a]

Average Slope	Characteristic Slope
Average Inclination	Ground Slope (Orthogonol)
Ground Slope (Valley-side)	Curvature of Slope Profile
Slopes greater than 50 Percent	

Relief Measurements[a]

Relief (Characteristic)	Average Elevation
Relief (Local)	Mean Valley Depth

(a) See Carr and Lopik (1962) for description of measurements.

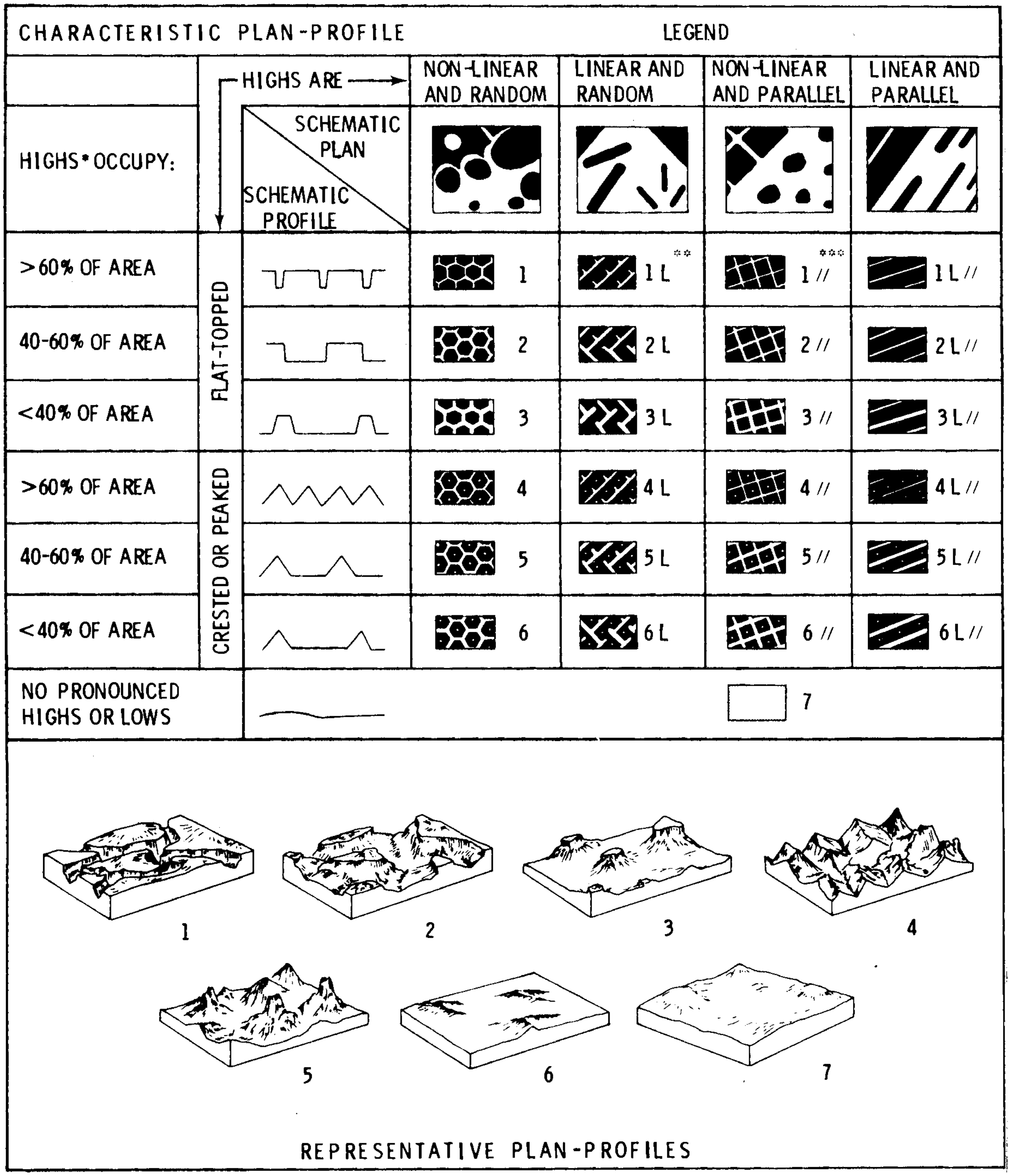

FIGURE I.3. Illustration of Characteristic and Representative Plan-Profiles (adapted from J. R. Van Lopik and C. R. Kobb 1958) terrain-induced airflow, transport and dispersion and also in presenting a useful perspective of the terrain. Obviously the accuracy of classifying the terrain this way depends upon the grid size and the area coverage, hence many details can be lost. Nevetheless, it is a technique that will probably see a lot of utilization in complex terrain studies in the future. A few examples are shown in Figure I.4.

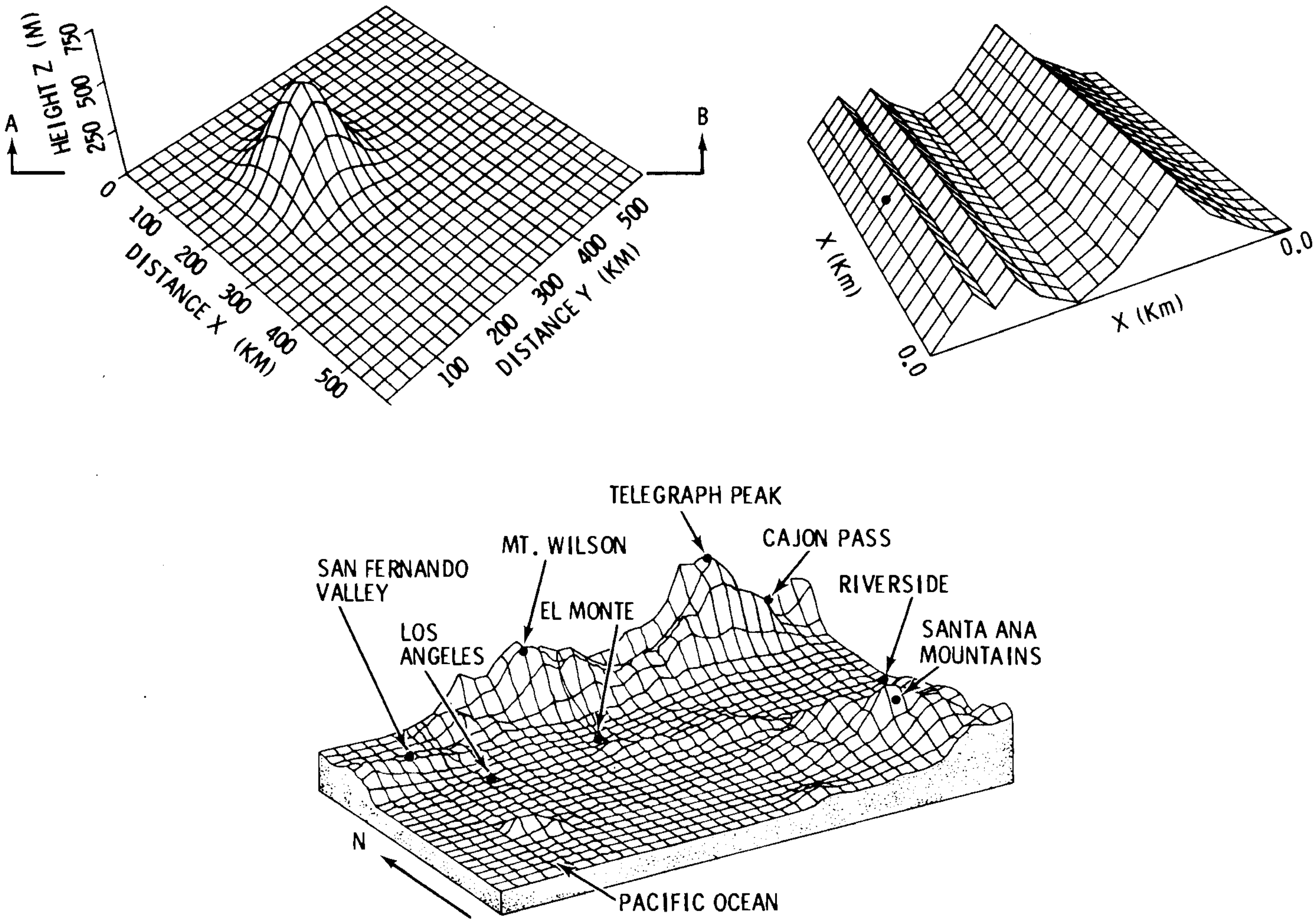

FIGURE I.4. Examples of Computer Generated Topography

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APPENDIX II

Known Geothermal Resource Areas of the Western United States

State	KGRA	Volcanic Cones	Type of Terrain
Arizona	Clifton		Valleys
	Gillard Hot Springs		
		San Francisco Mtns.	Older Volcanic Mtns.
California	Glass Mtn.		Volcanic Domes
	Lake City-Surprise Valley		N-S Desert Valley
	Lassen	Lassen Peak	Volcanic Domes
		Mt. Shasta	
	Wendel-Amedee		Basin (and lake)
	Beckworth Peak		Mountain and Basin
	Love Lady Ridge		Ridges and Canyons
	Little Horse Mtn.		Mountain Ridge
	Witter Springs		Wide Valley
	Geysers-Calistoga		Ridges, Valleys & Canyons
	Knoxville		Ridges and Canyons
	Mono-Long Valley		Leeside Basins or Valleys[(a)]
	Saline Valley		Leeside Desert Valley or Basin
	Coso Hot Springs		Leeside Mountain Slope
	Randsburg		Ridge and Desert Valley
	Sespe Hot Springs		Canyon
	Ford Dry Lake		Desert Basin (open)
	Salton Sea		Broad Desert Valley
	Brawley		" " "
	Glamis		" " "
	Dunes		" " "
	East Mesa		" " "
	Heber		" " "
Colorado	Alamosa County		Broad Mountain Basin (High Mountains Nearby)
	Valley View Hot Springs		North End of Mountain Basin (High Mountains Nearby)
	Mineral Hot Springs		North End of Mountain Basin (High Mountains Nearby)
	Poncha		High Mountain Valley
Idaho	Castle Creek		Dissected Sloping Terrain
	Bruneau		River Valley (Cliffs)
	Mountain Home		Desert Plain
	Crane Creek		Valley
	Vulcan Hot Springs		Mountains

(a) Leeside--East of the Sierra Nevada Mountains.

State	KGRA	Volcanic Cones	Type of Terrain
Idaho (cont)	Raft River		Desert Valley (High Mtns nearby)
	Island Park		Sloping Terrain
	Yellowstone		Sloping Terrain (Plateau)
Montana	Crowin Springs		Mountain Valley
	Boulder Hot Springs		Mountain Valley
	Marysville		Mountains
Nevada	Monte Neva		North-South Desert Valley
	Ruby Valley		Leeside of Mountain Ridge Shallow Valley
	Elko Hot Springs		Shallow Desert Valley
	Beowawe		NNE-SSW Desert Valley
	Hot Springs Point		NNE-SSW Desert Valley
	Rye Patch		North-South Desert Valley (Reservoir)
	Leach Hot Springs		NNW-SSE Desert Valley
	Kyle Hot Springs		North-South Desert Valley and Mountain Slope
	Dixie Valley		NNE-SSW Desert Valley
	Stillwater-Soda Lake		Desert Plain
	Salt Wells Basin		Small Open Basin
	Colado		Desert Valley & Mountain Slope
	Brady-Hazen		Desert Plain Isolated Peaks
	Wabuska		Desert Valley
	Wilson Hot Springs		Desert Valley
	Silver Peak		Mountain
	Darrough Hot Springs		NNE-SSW Desert Valley
	Warm Springs		Mountain Slope and Desert Valley
	Steamboat Springs		Valley (Leeside)[a]
	Moana Springs		Valley (Leeside)[a]
	San Emidio Desert		Desert Valley
	Gerlach		Black Rock Desert (Desert Valleys or Basins)
	Gerlach Northeast		Same as above
	Fly Ranch		Same as above
	Fly Ranch Northeast		Same as above
	Double Hot Springs		Same as above
	Trego		Same as above
	Soldier Meadow		Same as above
	Pinto Hot Springs		Same as above
	Baltazor		Desert Valley

(a) Leeside--East of the Sierra Nevada Mountains.

State	KGRA	Volcanic Cones	Type of Terrain
New Mexico	Baca (Valles Caldera)		Large Caldera (Mountain)
	San Ysidro		Canyon
	Socorro Peak		Small Mountain
	Lower Frisco Hot Springs		Canyon
	Gila Hot Springs		Canyons
	Radium Springs		Broad Desert Valley Isolated Mtn. Peaks
	Lightning Dock		Desert Valley
	Kibourne Hole		Gradual Sloping Desert Plain
Oregon	Vale Hot Springs		Wide Shallow Valley
	Alvrod		Narrow Desert Basin (Leeside of Ridge)
	Burns Butte		Sloping Desert Plain; Numerous Small Lakes
	Crump Geyser		Narrow Basin or Valley with Small Lakes
	Lakeview		Basin with Lake
	Klammath Falls		Broad Basin with Lakes
		Mt. McLoughlin	Volcanic Dome
	Summer Lake Hot Springs		Basin (Leeside of Ridge)
	Newberry Caldera		Large Caldera (Mountain)
	McCredie Hot Springs		Mountain Valley
		Mt. Jefferson	Volcanic Dome
	Belknap-Foley Hot Springs		Mountain Valley
		Three Sisters	Volcanic Domes
	Breitenbush Hot Springs		Mountain Valley
	Carey Hot Springs		Mountain Valley
	Mt. Hood	Mt. Hood	Isolated Volcanic Mountain
Utah	Monroe-Joseph		Mountain Valley
	CoveFort-Sulpherdale		Open Basin
	Roosevelt Hot Springs		Wide Desert Valley Ridges
	Thermo Hot Springs		Wide Desert Valley
	Lund		Desert Plain
	Newcastle		Desert Plain; Ridges
	Navajo Lake		Mountains
	Crater Springs		Desert Plain; Mountains Nearby
Washington	Indian Heaven		Mountain Slope
	Mt. St. Helens	Mt. St. Helens[a]	Isolated Volcanic Mountain or Dome
		Mt. Adams	Volcanic Dome
		Mt. Rainier	Volcanic Dome
	Kennedy Hot Springs		Mountain Valley
		Glacier Peak	
		Mt. Baker	Volcanic Domes

(a) Mt. St. Helens became active on March 20, 1980 and had a major eruption on May 18, 1980. A series of smaller eruptions have occurred since May 18.

APPENDIX III

Status of Coal Gasification and Liquefaction Projects

Status of Coal Gasification and Liquefaction Projects

A. COAL GASIFICATION PROJECTS

State	Companies	General Site	Process	Coal ton/Day	Remarks
Alabama	Tennesee Valley Authority	Muscle Shoals	Texaco's	168	Pilot Plant by 1980
California	Texaco, Inc. and Southern California Edison Co.	Near Daggett	Texaco's	1,000	Cool Water Coal Gasification Project, Plant Start-up 1983
Colorado	Cameron Engineers	Watkins	N.A.	34,250	Proposed Plant Operation by 1981
Kentucky	Texas Gas Transmission and Commonwealth of Kentucky	Western Kentucky near Ohio River	N.A.	N.A.	Pilot Plant by 1980, Commercial plant by 1983
	W. R. Grace & Co./DOE	N.A.	Texaco's	1,700	
	Pike County/State of Kentucky/DOE	Pike County (Eastern Kentucky)	Wellman-Galusha	36	Low-BTU Pilot Plant
Massachusetts	Massachusetts State and EG&G, Inc.	Fall River	N.A.	N.A.	Gas to run 1,000 MW power plant
Minnesota	Eric Mining Co. New Jersey Zinc Co./DOE	Hoyt Lakes	Woodall-Duckham	500	2 Low-BTU Pilot Plants
	Land O'Lakes/DOE	Perham	Wellman-Incandescent	72	Low-BTU Plant
Montana	Montana Power Co. Montana-Dakota Utilities Co.	Montana-Dakota Area	Hygas or similar	N.A.	
	Northern Natural Gas Co., Cities Service Gas Co.	Power River Basin	Lurgi Gasifier with Methanation	30,000 each	4 Plants Being Considered
	Washington Natural (Washington Energy Co.)	MaCone	N.A.	N.A.	

Status of Coal Gasification and Liquefaction Projects (con't.)

State	Companies	General Site	Process	Coal ton/Day	Remarks
New Mexico	El Paso Natural Gas Co.	Four Corners Area	Lurgi Gasifier with Methanation	28,250	Burnham Complex is on Navajo Indian Reservation
	WESCO (Western Gasification Co.) Texas Eastern Transmission Corp. Pacific Lighting Corp.	Four Corners Area (moved to new site in Wyoming)	Lurgi Gasifer with Methanation	28,625	On Navajo Indian Reservation (Canceled at New Mexico site)
North Dakota	American Natural Gas Co. (Michigan-Wisconsin Pipeline Co.) North American Coal Corp.	Beulah-Hazen Area	Lurgi Gasifer	N.A.	
	Natural Gas Pipeline Co. of America	Dunn County (Western North Dakota)	Lurgi Gasifier with Methanation	120,000	
	El Paso Natural Gas Co.	Southwestern North Dakota	N.A.	N.A.	Studying 4 plants; First to be in Operation in 1983
Pennsylvania	Koppers Co.	Verona	Koppers-Totzek	N.A.	Pilot Plant
	Bituminous Coal Research, Inc.	Homer City	Bi-Gas	120	
	U.S. Bureau of Mines	Bruceton	Synthane	70	
	Westinghouse Electric Corporation	Waltz Mill	Low-BTU Gas	N.A.	Commercial Size Plant to be Built at Terre Haute, Indiana
	Pennsylvania Gas & Water Co.	N.A.	HYGAS	5,000	
	New Jersey Zinc's	Palmerton	McDowell-Wellman	300	

Status of Coal Gasification and Liquefaction Projects (con't.)

State	Companies	General Site	Process	Coal ton/Day	Remarks
South Dakota	Michigan-Wisconsin Pipeline Co., American Natural Gas, Basin Electric Power Co.	Britton	N.A.	N.A.	
	Consolidation Coal Co.	Rapid City	CO_2 Acceptor	40	Pilot Plant
Tennessee	W.R. Grace & Co., Memphis Light, Gas and Water Division/DOE	Near Memphis	IGT U-Gas	2,800	Low-BTU Pilot Plant
Utah	Mountain Fuel Co	Emory County	Lurgi gasification with Methanation	N.A.	Planned Completion 1990
West Virginia	Columbia Gas System	N.A.	N.A.	N.A.	GGS Identifying Possible Sites
	Bureau of Mines, Consolidation Coal Co., Continental Oil Co.	Grants District of Wetzel County	In-Situ	N.A.	
Wyoming	Panhandle Eastern Pipeline Co., Peabody Coal Company	Eastern Wyoming	Lurgi Gasifer with Methanation	27,700	Plant Operation in Early 1980's
	Texaco, Inc., Natural Gas Pipeline of America, Montana-Dakota Utilities Co., Pacific Gas & Electric Company	Buffalo	N.A.	N.A.	Pilot Plant
	Carter Oil Co. (Exxon)	Northwest Wyoming	N.A.	N.A.	
	Cities Service-Northern Natural	Wyoming-Montana Border	N.A.	N.A.	

Status of Coal Gasification and Liquefaction Projects (con't.)

State	Companies	General Site	Process	Coal ton/Day	Remarks
Wyoming (Con't)	Bureau of Mines, Rocky Mountain Energy Co., Union Pacific Corp.	Hanna	In-Situ	N.A.	Since Late 1972
	Lawrence Livermore Lab. Gas Research Institute/ DOE	HOE Creek (N.E. Sector of Powder River Basin)	In-Situ	N.A.	Third Major Field Test

B. COAL LIQUEFACTION PROJECTS

State	Companies	General Site	Process	Coal ton/Day	Remarks
Kentucky	N.A.	Catlettsburg	H-Coal	600	
Montana	N.A.	Powder River (Belfry)	N.A.	N.A.	
	N.A.	Northern Great Plains (Circle West)	N.A.	N.A.	Two Plants Proposed
Washington	N.A.	Fort Lewis	Solvent Refining	N.A.	Pilot Plant
West Virginia	Fluor Engineers and Constructions	Cresap	N.A.	N.A.	Pilot Plant

APPENDIX IV

SOME DISPERSION EXPERIMENTS OR STUDIES IN COMPLEX TERRAIN

TABLE IV.1. Some Diffusion Experiments or Studies Involving Local Valley-Slope Wind Systems

Location	Source Type	No. of Releases	Tracer Type	Type of Topography	Reference
Columbia River Valley Trail, B.C., Canada	C-E-MP	---	SO_2	Complex Valley	Hewson and Gill (1944)
Cedar Mts. Dugway Proving Grounds, Utah	I-E-L	4	FP	Ridges and Canyons	McMullen and Perkins (1963)
Warm Springs Valley (Pyramid Lake), Nevada	C-S-P	27	OF UD FP	Ridge and Valley	Smith and Kauper (1963)
Tonopah, Nevada	I-E-L	7	FP	Ridge	Smith and Wolf (1963)
Twenty-Nine Palms, California	I-E-L	9	FP	Ridge	Smith (1965)
Johnstown, Pennsylvania	C-E-P	6	FP	Valley	Smith (1968)
Kanawha Valley, West Virginia	C-E, S-MP	---	---	Complex Valley	USNAPCA (1970)
Vandenberg AFB (Mt. Iron) California	C-S-P	113	FP	Ridges and Canyons	Hinds (and Nickola) (1970)
Parachute Creek and Roan Plateau, Colorado	C-S, E-P	17	FP SO_2 AgI	Canyon and Plateau	Battelle PNL (1972) Battelle PNL (1973)
Garmisch, Germany	C-S-P	3	FP	Valley and Hill	Reiter (1974) Reiter (1975)
Los Alamos and Parajito Canyons, New Mexico	C-S-P	7 (night) 6 (day)	FP	Sloping Canyons	Archuleta et al. (1978)

TABLE IV.1. (continued)

Location	Source Type	No. of Releases	Tracer Type	Type of Topography	Reference
Columbia River Valley Trail, B.C., Canada	C-E-MP	---	SO_2	Complex Valley	Whaley and Lee (1977)
Rio Blanco County Oil Tract C-a, Colorado	C-E-P	10	SF_6	Sloping Canyons	Bendel and Cresswell (1977)
Upper Hat Creek, B.C., Canada	C-E-P	14	OF SF_6	Complex Valley	Hovind et al. (1978)
Geysers Geothermal Area, California	C-E-P	3	FP	Valley and Ridges	Yohn (1977)
Geysers Geothermal Area, California	C-E-P	6 or more	CF_6 $CBrF_3$	Valley and Ridges	Huang and Baxter (1978)
Geysers Geothermal Area, California	C-E-P	---	H_2S	Valley and Ridges	Steffen, Wang, and Hidy (1978)
Geysers Geothermal Area, California	C-E-P	16	SF_6	Valley and Ridges	Smith, Knuth, and Giroux (1978)
Geysers Geothermal Area, California	C-E-P	10	SF_6 $SBrF_3$	Valley and Ridges	MRI (1978) and Knuth and Jensen (1978)
Geysers Geothermal Area, California	C-E-MP	---	H_2S	Valley and Ridges	Ruff (1979), and Cavanaugh and Ruff (1979)
Piceance Creek Valley, Colorado	C-E-P	2	SF_6	Valley	Chan (1979)
Coal Creek Valley, Colorado	C-E-P	---	SF_6	Valley	Nelson and Brown (1980)

TABLE IV.1. (continued)

Source Type		Tracer Type	
C	Continuous	FP	Fluorescent Particles
I	Instantaneous	OF	Oil Fog
P	Point	SO_2	Sulfur Dioxide
L	Line	AgI	Silver Iodide
E	Elevated	H_2S	Hydrogen Sulfide
S	Surface	SF_6	Silver Hexafluoride
MP	Multiple Point	$CBrF_3$	Fluorocarbons

TABLE IV.2. Some Diffusion Experiments in a Variety of Landforms that Measured Downwind Concentrations and/or $\sigma_y(x)$ and $\sigma z(x)$

Location	Type	Releases	Type	Topography	c(x)	$\sigma_y(x)$	$\sigma_z(x)$	Reference
Washington Nevada	I-E-L	7	FP	Rolling Hills, Ridge	x			Smith and Wolf 1963
Dugway Utah	I-E-L	4	FP	Desert, Mountain Ridge	x			Prophet 1963
Twenty-nine Palms, California	I-E-L	10	FP	Ridge	x	x	x	Smith 1965
Vandenburg AFB (Mt. Iron)	I-E-L	113	FP	Ridge-Canyon Foothills	x	x	x	Hinds 1970
Oceanside, California	I-E-L I-S-P C-S-P	55	FP	Shoreline, Rolling Hills	x	x	x	Smith and Nieman 1969
North Norway	C-S-P	53	FP OF	3 Deep Valleys, Fjord	x	x	s	Minott et al. 1977
Huntington Canyon, Utah	C-S-P C-E-P	6 5	SF_6 OF	Rugged Canyon	x			Start et al. 1975
Garfield, Utah	C-E-P C-S-P	8 2	SF_6	Mountain Ridge	x			Start et al. 1974
Julich, F. R. Germany	C-E-P	44[a]	$CuSO_4$	Rhine River Valley z_0 = 0.4 to 1.0 m	x	(b)		Vogt et al. 1974, Draxler 1979, and Miller 1978
Karlsruhe, F. R. Germany	C-E-P	25[a]	HTO HC	Rhine River Valley z_0 = 1.1 m	x	(b)		Thomas et al. 1974, Draxler 1979, and Miller 1978
Loisach Valley F. R. Germany	C-E-P C-E-P	11	FP	Deep Valley, Alp Foothills	x	x		Reiter 1974 and 1975, and Draxler 1979
Widow's Creek Tennessee River Valley	C-E-MP	---	SO_2	Shallow Valley		x	x	Hanna 1980

(a) $CuSO_4$--Copper Sulfate; HTO--Tritiated Water Vapor; Hc--Hydrocarbons.
(b) Ground level air concentrations were used to calculate $\sigma_y(x)$ and $\sigma_z(x)$ by assuming a Gaussian plume.

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