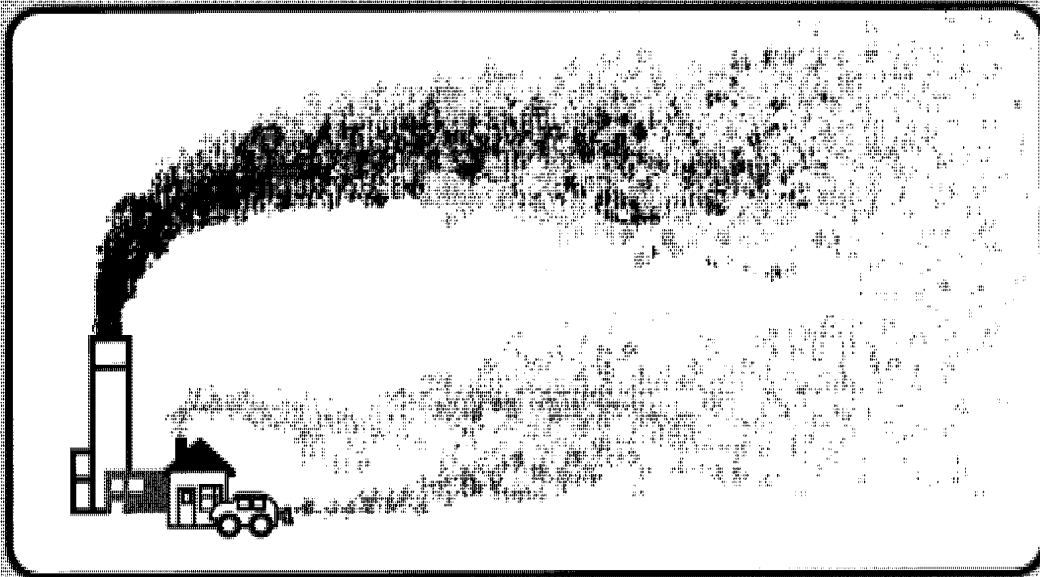


PLUME DISPERSION IN ALBERTA



Alberta
ENVIRONMENT

**PLUME DISPERSION
IN
ALBERTA**

by

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PREFACE

This compendium is the outcome of two separate, but related activities in which the authors have been engaged for more than a decade. Reviews of world literature and local studies of diffusion were undertaken prior to issuing Alberta Environment's first modelling guideline in 1978 and releasing a new draft for public comment in 1989. Over the years, the authors often taught short courses on dispersion meteorology and modelling, mainly for professionals in other scientific and technical specializations.

In this volume we have combined our educational notes about the fundamentals of our science with the results of ongoing review of the scientific literature. The purpose of this document is to summarize current information about plume dispersion in the province of Alberta, within the context of general scientific understanding of the lower atmosphere. Each topic is introduced with an explanation of concepts and a definition of technical terms. The review encompasses virtually all published Alberta work up to the end of 1988; some more recent work was added during the final editing. Alberta references are always cited directly in the text; world literature references are cited in parenthesis following a statement of generally accepted scientific understanding. Emphasis is given to topics of local relevance.

Meteorology is the science dealing with the phenomena of the atmosphere. This includes not only the physics, but the history, chemistry, geology and biology of the atmosphere and many of the direct effects of the atmosphere upon the earth's surface, the oceans and life in general. The study of atmospheric processes governing the behavior and fate of airborne chemicals constitutes a specialization called *air quality meteorology*. The atmosphere may be viewed in different ways: as a medium for various physical processes, as a carrier for chemical, geological or biological substances, or as a setting in which a number of geophysical, chemical and biological processes take place.

This review is concerned primarily with the continuous emission of chemicals from exhaust stacks (or chimneys as they are sometimes called). When the material leaving a stack is visible it assumes a distinctive featherlike form called a *plume*. By extension the word plume is also used to describe invisible emissions since measurements can still indicate the presence of the structure. The report does not deal with accidental releases of toxic substances, urban air quality, dense gases, fog interactions, acid rain or the greenhouse effect. Its focus is on the ground-level concentrations that result from plumes in the atmosphere.

Diffusion is defined as the exchange of air parcels and their conservative properties (such as heat, moisture, momentum) between neighboring regions of the atmosphere. The rate of transfer or mixing is several orders of magnitude greater than that due to molecular agitation. To be rigorous about the terminology one should probably refer to *turbulent diffusion* or *eddy diffusion* as distinct from *molecular diffusion*. The above definition places no restriction on the scale of the fluid motions involved in the exchange. Therefore, it can include the process commonly called *transport*, which is defined as the movement of identifiable air parcels from one place to another in an organized manner that can be treated by the equations of motion. However, when the term *transport* is used in this manner, the term *diffusion* is usually restricted to those apparently random motions of a scale too small to be treated by the equations. The dictionary defines *dispersion* as the act or process of breaking up and going in different directions, spreading or distributing irregularly, scattering in different directions. Thus *dispersion* is a term even broader than diffusion and includes all processes that affect the fate of an airborne chemical, such as chemical transformation, rainout and surface removal.

The primary elements of air quality meteorology are three: (1) wind; (2) temperature stratification; and (3) turbulence. These are introduced in the first chapter; descriptions of measurement methods and data display techniques may be found in the appendices. The three elements are integrated in the discussion of the boundary layer, and the rise of a plume into the atmosphere. This leads directly to atmospheric diffusion over gentle terrain and a brief description of the removal processes. Finally, the complex subject of flow and diffusion over rugged terrain is surveyed. Each chapter has a separate reference list and a separate summary section so that it can be read somewhat independently. The division of labor between authors gave principal authorship of chapters 1, 2 and 4 to R.P. Angle and of chapters 3 and 5 to S.K. Sakiyama.

CONTENTS

	Page
PREFACE	i
TABLES	v
FIGURES	viii
ACKNOWLEDGEMENTS	xiii
 1. ELEMENTS OF DIFFUSION METEOROLOGY	 1-1
1.1 Wind	1-1
1.2 Temperature Stratification	1-22
1.3 Turbulence	1-39
1.4 Weather Systems	1-78
1.5 Summary	1-99
1.6 References	1-101
 2. THE ATMOSPHERIC BOUNDARY LAYER	 2-1
2.1 The Neutral Boundary Layer	2-1
2.2 The Unstable Boundary Layer	2-6
2.3 The Stable Boundary Layer.. . . .	2-13
2.4 The Thickness of the Planetary Boundary Layer.	2-19
2.5 Power Law Forms of the Wind Profile.	2-26
2.6 Boundary Layer Evolution	2-28
2.7 Plume Behavior	2-40
2.8 Summary	2-54
2.9 References	2-55
 3. PLUME RISE	 3-1
3.1 General Considerations	3-1
3.2 Plume Rise in Neutral and Unstable Conditions	3-4
3.3 Plume Rise in Stable Conditions	3-8
3.4 Final Rise in Neutral and Unstable Conditions	3-11
3.5 Final Rise in Stable Conditions.	3-15
3.6 Plume Penetration of Elevated Inversions	3-15
3.7 Stack Aerodynamic Downwash	3-18
3.8 Behavior of Moist Plumes	3-21
3.9 Plume Rise from Gas Turbines and Reciprocating Engines	3-22
3.10 Summary	3-23
3.11 References	3-24

CONTENTS

	Page
4. ATMOSPHERIC DISPERSION OVER GENTLE TERRAIN	4-1
4.1 The Gaussian Plume Model	4-2
4.2 Empirical Systems for Plume Spreads	4-9
4.3 Plume Spreads from Taylor's Theory	4-24
4.4 Plume Spreads from Interrelated Theories	4-29
4.5 Averaging Time	4-34
4.6 Buoyancy	4-43
4.7 Plume Height	4-45
4.8 Longer Travel Time	4-48
4.9 Wind Shear	4-50
4.10 Surface Roughness	4-51
4.11 Along-wind Spread	4-52
4.12 Nonstationarity and Calms	4-53
4.13 Concentration Fluctuations	4-55
4.14 Chemical Transformations	4-59
4.15 Removal.	4-61
4.16 Summary	4-64
4.17 References	4-66
5. FLOW AND DIFFUSION OVER RUGGED TERRAIN	5-1
5.1 Wind and Turbulence Effects	5-1
5.2 Mathematical Modelling of Flow over Ridges and Hills	5-21
5.3 Mathematical Diffusion Modelling Over Rugged Terrain	5-24
5.4 The Effect of Buildings on Plume Dispersion.	5-42
5.5 Physical Modelling of Transport and Diffusion Over Complex Terrain.	5-55
5.6 Summary	5-61
5.7 References	5-63
APPENDIX A. MEASUREMENT TECHNIQUES	
A.1 Wind	A-1
A.2 Temperature Stratification	A-10
A.2 Turbulence	A-12
A.4 Boundary Layer Height.	A-20
A.5 Dilution.	A-22
A.6 References	A-24
APPENDIX B. DATA PRESENTATION TECHNIQUES	
B.1 Wind Direction	B-1
B.2 Wind Speed	B-5
B.3 Wind Vector (Speed and Direction).	B-5
B.4 References	B-18

TABLES

	Page
1.1 Averages of inversion types for Edmonton	1-38
1.2 Joint frequency distribution of temperature gradient with wind direction for Edmonton CBC Tower	1-42
1.3 Joint frequency distribution of temperature gradient with wind direction for Calgary Bonnybrook Tower	1-43
1.4 The 18-m to 8-m temperature gradient at Ellerslie	1-44
1.5 Summary of inversion characteristics measured by minisonde near Edmonton 1977 - 1982	1-45
1.6 Key to stability categories	1-66
1.7 Frequency of occurrence of Pasquill stability classes	1-67
1.8 Stability classification using net radiation	1-71
1.9 Turbulence classification based on wind fluctuations	1-72
1.10 Bulk Richardson numbers measured on 91-m towers in Calgary and Edmonton.	1-74
1.11 Relation among turbulence typing methods.	1-76
1.12 Relation between Pasquill class, Monin-Obukhov length and roughness .	1-75
1.13 Values of Kazanski-Monin parameter corresponding to Pasquill classes .	1-77
1.14 Airmasses and fronts in the conventional Canadian classification	1-80
2.1 Summary of experimentally determined roughness lengths	2-4
2.2 Summary of equations describing the neutral atmospheric boundary. . .	2-7
2.3 Summary of equations describing the unstable atmospheric boundary . .	2-9
2.4 Summary of equations describing the stable atmospheric boundary . . .	2-16
2.5 Mean maximum (afternoon) mixing heights at Edmonton	2-21

TABLES

		Page
2.6	Median values of kink mixing height at Ellerslie	2-24
2.7	Alternative definitions of the thickness of the atmospheric boundary layer	2-27
2.8	Summary of power law exponents observed in Alberta	2-29
2.9	Seasonal hourly values of mixing height at Fort McMurray	2-36
2.10	Types of plume behavior and associated meteorology	2-44
2.11	Monthly distribution of stability at Calgary.	2-53
3.1	Observational studies in Alberta of buoyant plume rise in neutral/unstable conditions	3-6
3.2	Observational studies in Alberta of buoyant plume rise in stable conditions.	3-10
3.3	Final rise formulations	3-13
3.4	Observations in Alberta of final rise in neutral/unstable conditions	3-16
4.1	Current working theories of diffusion.. . . .	4-3
4.2	Summary of basis and scope of original Pasquill-Gifford curves	4-14
4.3	Single power law approximations to the Pasquill-Gifford curves	4-15
4.4	BNL plume standard deviations.	4-19
4.5	TVA diffusion coefficients	4-19
4.6	Briggs plume spread formulas for elevated small releases	4-22
4.7	Summary of Draxler's equations	4-27
4.8	Other universal functions	4-28
4.9	Power law approximations of vertical spread	4-30
4.10	Sigma values from neutral to unstable using convective scaling	4-33

TABLES

	Page
4.11 Classification of turbulent diffusion from a fixed point source	4-37
4.12 Summary of approaches to concentration fluctuations.	4-56
5.1 Observation of up- and down-valley flow in Alberta	5-12
5.2 Nocturnal valley flow characteristics in Alberta	5-14
5.3 Summary of wind field models	5-24
5.4 Wind field models as applied in Alberta	5-26
5.5 Observational studies in Alberta on the applicability of potential flow	5-28
5.6 Plume spread observations in Alberta in rugged terrain	5-36
5.7 Numerical model applications for complex terrain in Alberta.	5-44
A.1 Anemometers suitable for diffusion applications	A-3
A.2 Remote sensors for winds above the surface layer	A-9
A.3 Methods for measuring temperature stratification	A-11
A.4 Summary of turbulence data from aircraft flights in Fort McMurray area	A-19
A.5 Methods of measuring the thickness of the boundary layer.	A-21
A.6 Equipment and design characteristics for SF ₆ studies.	A-23
B.1 Usual method of converting from tens of degrees.	B-2
B.2 Debiasing procedure for direction frequencies.	B-3
B.3 Beaufort wind speed equivalents	B-6
B.4 Normalized radius of the percentile circle in standard vector deviation roses.	B-13

FIGURES

	Page
1.1 Mean seasonal (monthly) sea level pressure in kilopascals	1-3
1.2 Balance between pressure gradient force and Coriolis force	1-5
1.3 Balance between pressure gradient force, Coriolis force and centripetal force	1-5
1.4 Balance of forces with friction added.	1-7
1.5 The Ekman spiral of wind with height.	1-7
1.6 Wind roses for selected Alberta stations	1-8
1.7 Diurnal variations of 10-m and 91-m winds in Calgary	1-11
1.8 Percentage of wind speed above a specified value	1-14
1.9 Return periods for extreme wind speeds, Medicine Hat	1-15
1.10 Occurrences of persistent light winds lasting 24 to 47 h.	1-16
1.11 Occurrences of persistent light winds lasting 48 h and longer	1-17
1.12 Duration of light winds in Edmonton and Calgary	1-18
1.13 Vector mean winds over Alberta at the 85-kPa level	1-20
1.14 Standard vector deviation of winds over Alberta at the 85-kPa level	1-21
1.15 Trajectories of air parcels leaving the Alberta oil sands area	1-23
1.16 Trajectories of air parcels entering the Alberta oil sands area	1-23
1.17 Frequencies of 72-h air parcel motion out of the Alberta oil sands area at the 85-kPa level	1-24
1.18 Frequencies of 72-h air parcel motion into the Alberta oil sands area at the 85-kPa level	1-25
1.19 Frequency of ground-based inversion at 05:00 MST	1-34
1.20 Frequency of ground-based inversion at 17:00 MST	1-35

FIGURES

	Page
1.21 Seasonal frequency and intensity of ground-based inversions at Edmonton..	1-36
1.22 Inversion types..	1-37
1.23 Cumulative frequency distribution of temperature gradient	1-39
1.24 Diurnal variation of temperature gradient	1-41
1.25 The energy cascade	1-51
1.26 Spectrum of turbulence.	1-51
1.27 Autocorrelation curve.	1-55
1.28 Conventional presentation of vertical wind spectra.	1-55
1.29 Wind spectra at Suffield, Alberta.	1-58
1.30 Solar radiation by month, hour and cloud cover	1-69
1.31 Stability number P from heat flux and wind speed	1-70
1.32 Meteorological scale definitions	1-79
1.33 Arctic front location.	1-82
1.34 Streamlines and zones of confluence.	1-84
1.35 Airmass frequencies	1-85
1.36 Number of days with lows	1-87
1.37 Number of different lows	1-88
1.38 Tracks of lows	1-89
1.39 Tracks of highs.	1-90
1.40 Number of days with highs	1-91
1.41 Number of different highs.	1-92
1.42 Most frequent Alberta weather map types.	1-94
1.43 Schematic of mountain wind systems	1-96

FIGURES

	Page
2.1 Schematic of the neutral PBL	2-2
2.2 Schematic of the unstable PBL.	2-10
2.3 Scaling regions for the unstable and stable PBL.. . . .	2-11
2.4 Schematic of the stable PBL.	2-14
2.5 Schematic of nocturnal jet formation	2-18
2.6 Mean maximum afternoon mixing heights.	2-22
2.7 Hourly median mixing heights at Fort McMurray	2-25
2.8 Mean breakup and onset times of nocturnal inversions	2-34
2.9 Diurnal variations of P stability and mixing heights	2-38
2.10 Mean daily global radiation	2-39
2.11 Plume behavior types	2-41
3.1 Depiction of turbulent shear zone and entrainment.	3-3
3.2 Schematic illustrating basic definitions for plume rise	3-3
3.3 Depiction of plume rise in a stable atmosphere	3-9
3.4 Plume penetration of elevated inversions	3-17
3.5 Depiction of stack aerodynamic downwash.	3-20
3.6 Modified speed ratio versus Froude number.	3-20
4.1 Vertical dispersion coefficient as a function of downwind distance	4-10
4.2 Horizontal dispersion coefficient as a function of downwind distance	4-11
4.3 Dispersion curves of the Brookhaven National Laboratory compared with the Pasquill-Gifford curves	4-18
4.4 Briggs rural dispersion coefficients compared to Pasquill-Gifford	4-23

FIGURES

	Page
4.5 Smith's vertical spread compared to Pasquill-Gifford	4-31
4.6 Plumes classified by observation time.	4-36
4.7 Schematic representation of a quasi-steady plume	4-36
5.1 Channel and gap winds.	5-2
5.2 Lee waves, cloud and rotor flow	5-5
5.3 Hydraulic jump.	5-5
5.4 Flow near terrain obstacles.	5-6
5.5 Katabatic and anabatic flows	5-9
5.6 Valley circulations	5-10
5.7 Hourly percent frequency of wind directions in Canyon Creek valley . .	5-10
5.8 Wind speed and direction profiles in Canyon Creek valley.	5-13
5.9 Time series of potential temperature profiles	5-17
5.10 Airflow toward a steep bluff	5-21
5.11 Separation in strong cross-valley winds	5-21
5.12 Potential flow streamlines past a hemisphere	5-29
5.13 Various plume height adjustment approaches	5-34
5.14 A stable flow streamline across a terrain obstacle	5-38
5.15 Idealized stratified flow about hills	5-38
5.16 Cooling tower plume outline and centreline trajectory	5-40
5.17 Dimensionless elevation as a function of wind speed	5-40
5.18 Behavior of plumes emitted directly above a building	5-45
5.19 Flow pattern around a rectangular block	5-46

FIGURES

	Page
5.20 Vortex and venturi effects for building clusters	5-49
5.21 Definition and calculation of minimum descent height	5-51
5.22 University of Alberta small wind tunnel	5-60
A.1 Atmospheric Environment Service weather stations in and near Alberta	A-4
A.2 Upper air stations operated by the Atmospheric Environment service in western and northwestern Canada	A-8
B.1 Different types of wind roses for Red Deer	B-4
B.2 Weibull probability paper.	B-7
B.3 Wind rose with speed segments.	B-8
B.4 Speed-direction frequency diagram.. . . .	B-10
B.5 Box plots for wind vector information.	B-11
B.6 Extracting information from standard vector deviation wind roses.	B-14
B.7 Polar isopleth diagram	B-15

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1. **ELEMENTS OF PLUME DISPERSION METEOROLOGY**

One of the main objectives of plume dispersion meteorology is the prediction of future ground-level concentrations resulting from present or proposed emissions of chemicals to the atmosphere. Ideally, such "forecasts" would include the frequency distribution of ground-level concentration averaged over time periods consistent with those specified in ambient air quality objectives, and these distributions would be produced as functions of time of day and season for all locations of concern. In practice, the statistics of past meteorological conditions must be used under the assumption that they will be reasonable estimators of future weather. Ground-level concentrations are primarily controlled by three meteorological elements – wind, temperature stratification, turbulence – and the various scales of temporal and spatial changes in these elements.

1.1 **WIND**

Wind is the term used to describe air in motion. The motion is three-dimensional and the equivalent term "air velocity" indicates that, mathematically speaking, wind is a vector, that is, a quantity having both magnitude and direction. Usually the vertical component is disregarded and wind is considered to be a two-dimensional vector specified by the horizontal wind speed and direction. By convention the wind direction is identified as the direction *from* which the wind blows, and it is measured clockwise from true north. Measurement considerations and techniques are discussed in Appendix A.

Wind direction determines where a plume will be carried and which receptors may potentially be affected. Wind speed determines the travel time from the source to the receptor and also provides a measure of the volume of air available for dilution. If one visualizes "boxes" of air of unit dimension the wind speed indicates how many such boxes pass the stack in unit time. If, for example, the wind speed is doubled and the mass of chemical emitted in unit time is constant, each box will then contain only one half as much material as before. The concentration is reduced through "stretching" in the forward direction. Wind speed is also a determinant of the height to which a plume will rise after being emitted from a stack.

In essence the wind blows because of unequal heating of the earth's surface by the sun. Air moves in order to transport heat energy from regions of surplus to regions of deficit. On a global scale, energy is transported poleward in response to the temperature difference between the tropics and the poles. Thus it is often said that the atmosphere acts like a gigantic heat engine.

1.1.1 Horizontal Forces

Differential heating produces variations in air pressure with distance. Such changes in space mean that air parcels are subject to a net force which tends to drive them from the region of high pressure to the region of low pressure. Mathematically this pressure gradient force is given by

$$PGF = -1/\rho \partial P/\partial n$$

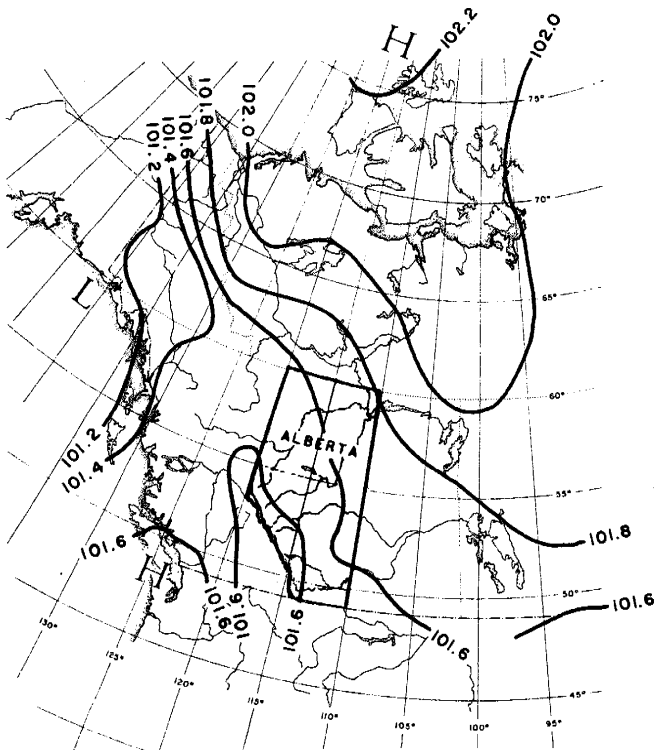
where P is the pressure, n is the distance normal to the isobars, and ρ is the air density. The closer the isobar spacing on a surface weather chart, the stronger the gradient and the stronger the winds. Horizontal pressure differences are usually of the order of one kilopascal per thousand kilometres (Figure 1.1). In central Alberta mean annual air density is about 1.2 kg/m^3 ($P = 93 \text{ kPa}$, $T = 1^\circ\text{C}$), so the pressure gradient force per unit mass is typically $8 \times 10^{-4} \text{ N/kg}$.

Once in motion under the influence of the pressure gradient force, an air parcel does not move in the straight line that one might expect. Instead it will curve to the right in the northern hemisphere. The deflection is caused by the rotation of the earth. Although the air parcels actually do move in straight lines, to the observer on the rotating earth they appear to behave as if a "deflecting force" were acting upon them. This apparent deflecting force is called the *Coriolis force* and it is given by

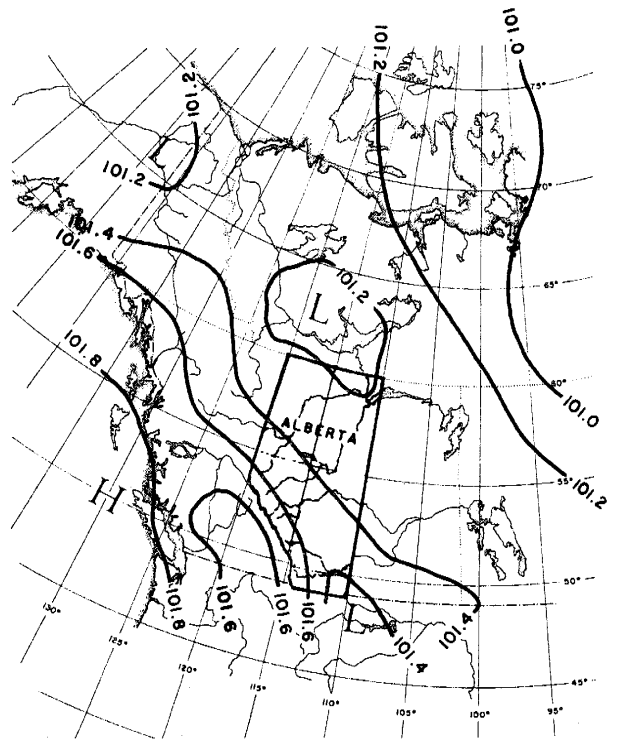
$$CF = 2\omega V \sin \phi$$

where ω is the angular velocity of the earth's rotation, V is the speed of the air parcel and ϕ is the latitude.

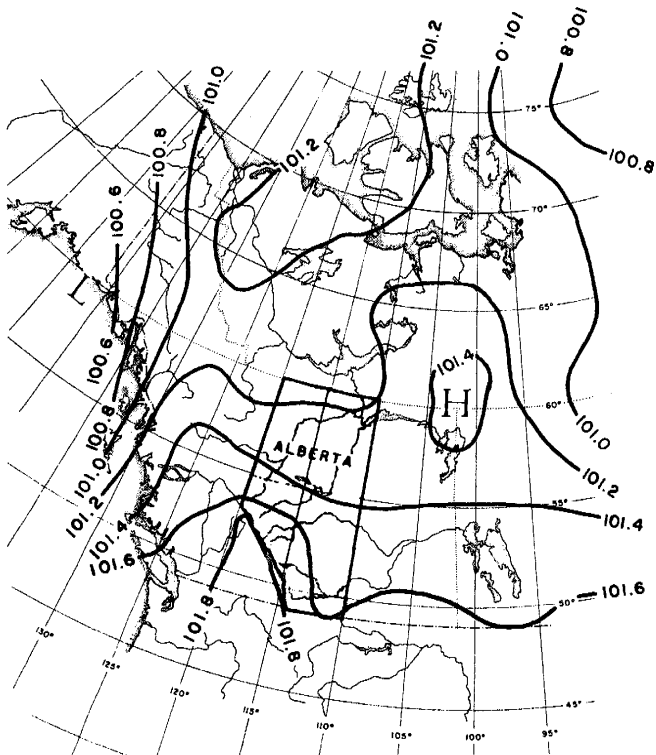
The magnitude of the force depends directly upon the speed of the moving air parcel. In central Alberta ($\phi = 54^\circ$) an air parcel moving at 10 m/s will experience a



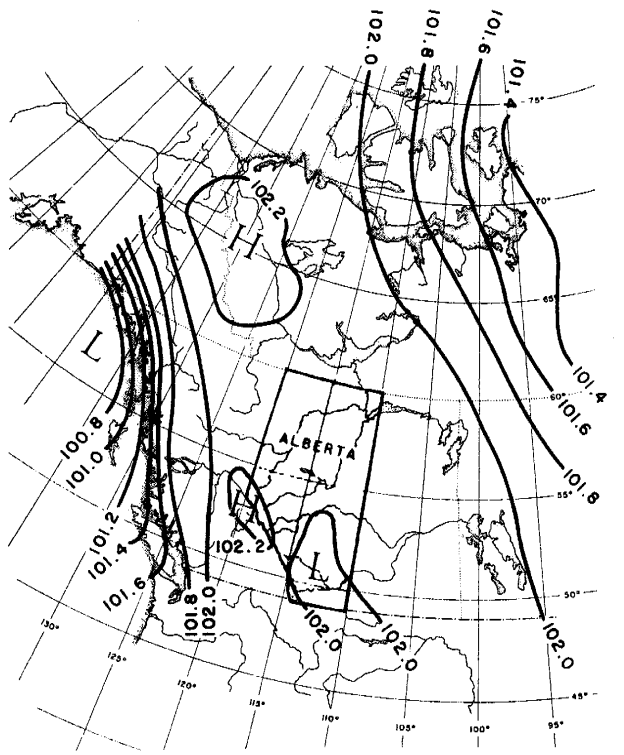
Spring (April)



Summer (July)



Autumn (October)



Winter (January)

Figure 1.1 Mean seasonal (monthly) sea level pressure in kilopascals.

Coriolis force per unit mass of 1.2×10^{-3} N/kg directed to the right of its direction of travel.

An air parcel initially at rest will first be accelerated by the pressure gradient force. As it begins to move, the Coriolis force pushes it to the right. The resultant of the two forces is then not as great as that of the PGF alone and the rate of speed increase will be reduced somewhat. The changes in direction and speed continue until the resultant force on the air parcel is zero and it travels at constant speed parallel to the isobars (Figure 1.2). From the balance of forces, this speed will be given by

$$V_g = -PGF/2\omega\sin\phi$$

and is known as the *geostrophic wind*. For central Alberta, with the order of magnitude pressure gradient given earlier, the geostrophic wind would be about 7 m/s.

The magnitude geostrophic wind at a given level is proportional to the pressure gradient at that level. The difference between geostrophic winds at two levels is called the *thermal wind*. The thermal wind "blows" parallel to isotherms with warm air on the right. Its strength is proportional to the horizontal temperature gradient.

If the isobars exhibit significant curvature as in the vicinity of pressure centres, then consideration must also be given to the centripetal force, given by

$$F_c = -V^2/R$$

where V is the wind speed and R is the radius of the circular trajectory.

Near a low pressure centre the centripetal force acts in the same direction as the pressure gradient force, resulting in a wind somewhat less than geostrophic, whereas near a high pressure centre it acts in the same direction as the Coriolis force, producing a wind somewhat greater than geostrophic (Figure 1.3). The air motion that occurs as a result of the balance of these three forces is called the *gradient wind*. The centripetal force is generally small; except in systems with high winds in tightly curved paths, the difference between gradient and geostrophic winds is insignificant.

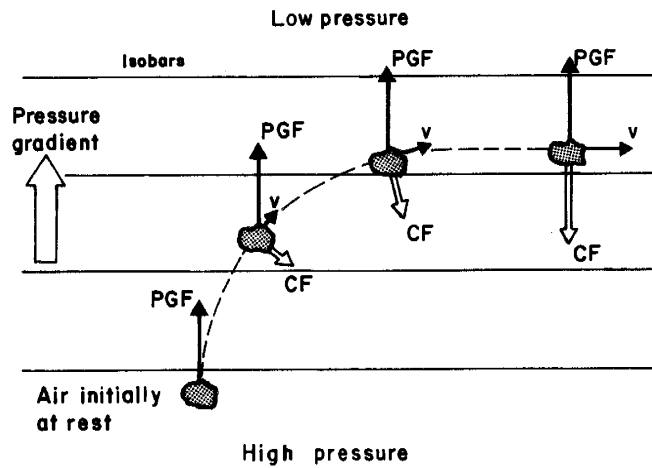
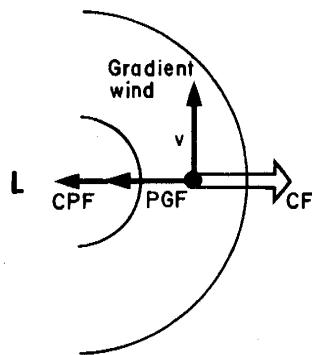
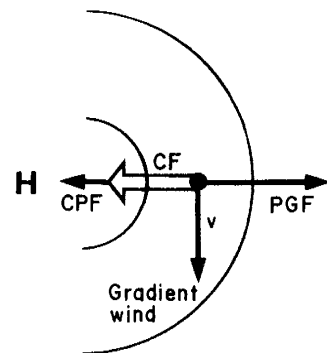


Figure 1.2 Balance between pressure gradient force and Coriolis force. Airmass responding to a pressure gradient force PGF is imagined to accelerate initially from rest. Once it gains velocity v , the Coriolis force CF deflects it until a force balance is reached in geostrophic flow.



(a) Low pressure



(b) High pressure

Figure 1.3 Balance between pressure gradient force PGF, Coriolis force CF, and centripetal force CPF, produces the gradient wind.

Friction is the name given to the force brought about by the exchange of momentum between the atmosphere and the earth's surface. Since the earth's surface cannot move, the net effect of the momentum exchange is to decrease the speed of the wind. If a balance of forces is established with friction acting opposite the direction of movement of the air parcel, then the wind will be turned towards low pressure (Figure 1.4). Alternatively, one might say that friction slows the wind and decreases the Coriolis force, allowing the pressure gradient force to move the air parcel across the isobars. The angle between the resulting wind direction and the isobars increases as the frictional force increases. Over relatively smooth surfaces like the oceans or grasslands, the angle will be 10 to 20° while over very rough surfaces like cities the angle may be 30 to 50°. The corresponding wind speeds measured at a height of 10 m above the ground will be reduced to about 0.9 and 0.5, respectively, of the geostrophic value. Both the angle and speed reduction are also strongly dependent on temperature stratification. (for example, Moses and Hess 1976).

The retarding effect of the earth's surface is transferred upward by eddy viscosity or turbulence. The further away from the earth's surface, the smaller the friction force. The balance of forces requires the wind direction to change clockwise with height, producing the Ekman spiral (Figure 1.5). At the top of the spiral the wind will be geostrophic. The layer in which the effects of friction are present is called the friction layer. It is generally in the order of one kilometre thick but this depends upon terrain roughness, temperature stratification, pressure gradient and other features of the large-scale flow. Wind direction and speed are usually measured at 10 m above the ground (Appendix A).

1.1.2 Wind Direction

Diagrams representing the frequencies of winds from various directions at a fixed location are known as *wind roses*. The most common form consists of a small circle from which eight or sixteen radii emanate, one for each compass point. The length of each line is proportional to the frequency of wind from that direction. The longest line, corresponding to the most frequent direction, gives the *prevailing wind*. Wind roses for a number of Alberta stations are given in Figure 1.6. Other types of wind roses are discussed in Appendix B.

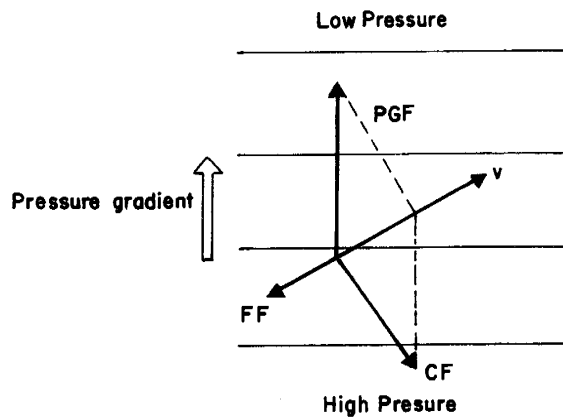


Figure 1.4 Balance between pressure gradient force PGF , Coriolis force CF and the frictional force FF (which must be directed opposite to the wind velocity v).

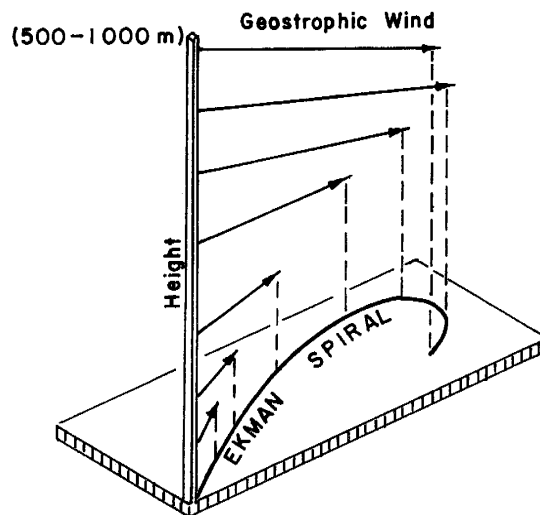


Figure 1.5 The Ekman spiral of wind with height in the northern hemisphere. The wind attains the geostrophic velocity at between 500 and 1 000 metres in the middle and higher latitudes as frictional effects become negligible.

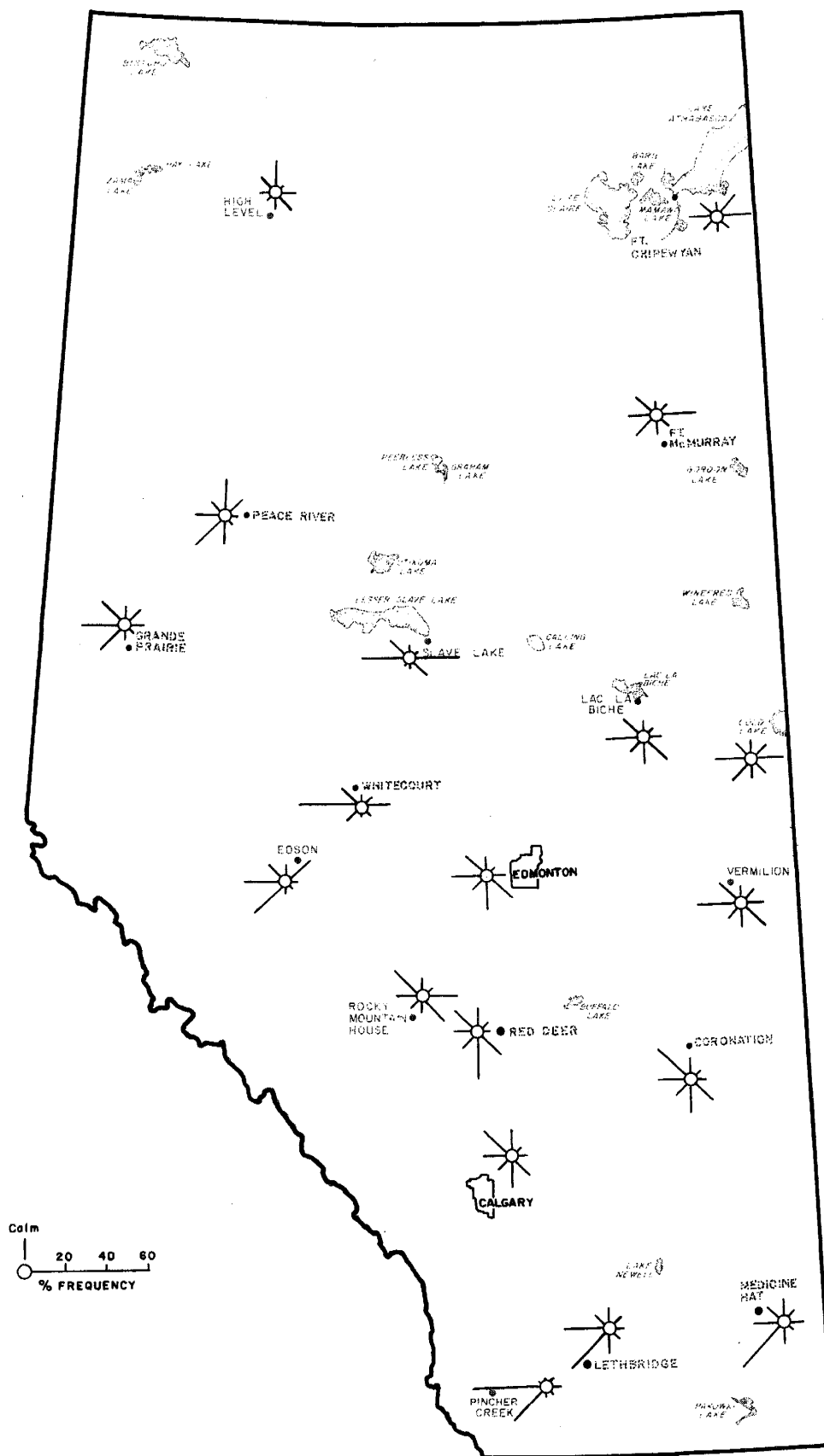


Figure 1.6 Wind roses for selected Alberta stations 1955-1980.

Care must be exercised in making decisions based on the idea of a single prevailing wind direction. One or more other directions may be almost as frequent. When these secondary directions are removed from the primary direction by several compass points, ignoring them could lead to serious errors. Because chemical concentrations are dependent upon temperature stratification and turbulence, as well as wind speed, the direction of greatest plume impact can only be ascertained by examination of joint frequency distribution (Appendix B).

Some measure of the variability of the wind direction over periods on the order of a day is often useful in characterizing various types of pollution events and the potential for such events. *Wind direction persistence* refers to the tendency of the wind to remain in the same sector for consecutive time intervals. One type of persistence diagram gives the probability that the wind, once having turned into a given sector, remains in that sector for consecutive time intervals of up to 100 hours (Slade 1968). With direction data for 16 compass points each sector will be 22.5° in width. Similar diagrams can also show the probability that the wind, once having turned into a given sector, remains within ± 1 or ± 2 compass points.

Persistence frequency has been found to obey a log-normal distribution at two of three stations in India (Shirvaikar 1972). The return period for direction persistences equal to or greater than T can be calculated from

$$R(T) = 1/(1-F(T)) \left(\sum_i^{T-1} f_i \right) (1+C/100)$$

where $F(T)$ is the probability of occurrence of periods of less than T, f_i is the frequency of persistence period i, and C is the percentage of calms.

The sector technique has a shortcoming in that the switching of the wind from the specified sector for a single observation will end the period of persistence even though the wind may resume the original direction immediately thereafter. The range of wind directions and the standard deviation of wind directions in a given period are logical parameters to employ. Small variability means that a plume will be present over a particular downwind area for extended periods, leading to relatively high average concentrations. The term "steady wind incident" (SWI) describes such situations which

could potentially result in local exceedances of ambient air quality objectives (Gorr and Dunlap 1977). On the other hand, large variabilities are often associated with periods of atmospheric stagnation which lead to poor air quality in urban centres where there are a multitude of small pollutant sources.

A *consistency index* can be defined (Heidorn 1978) as

$$CI = (100/H) (C_n^2 + C_e^2)^{1/2}$$

$$\text{where } C_n = \sum H_i \cos \theta_i$$

$$C_e = \sum H_i \sin \theta_i$$

$$H = \sum H_i$$

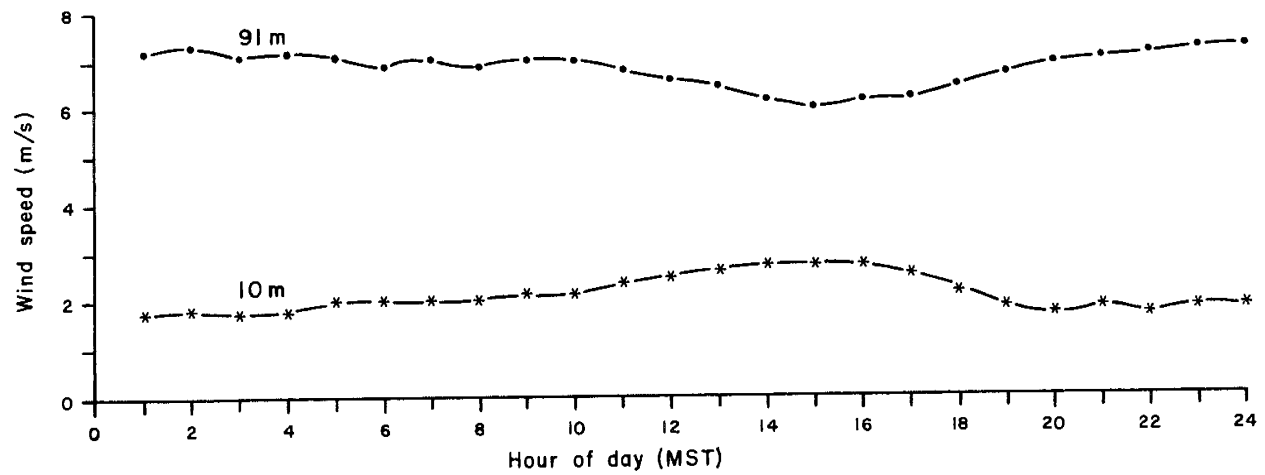
and H_i is the number of hours that the wind blows from direction category i of which θ_i is the mid-angle. This index has the value zero when the sum of the components in the N-S (C_n) and E-W (C_e) planes is zero and 100 when the direction is constant throughout the period.

1.1.3 Wind Speed

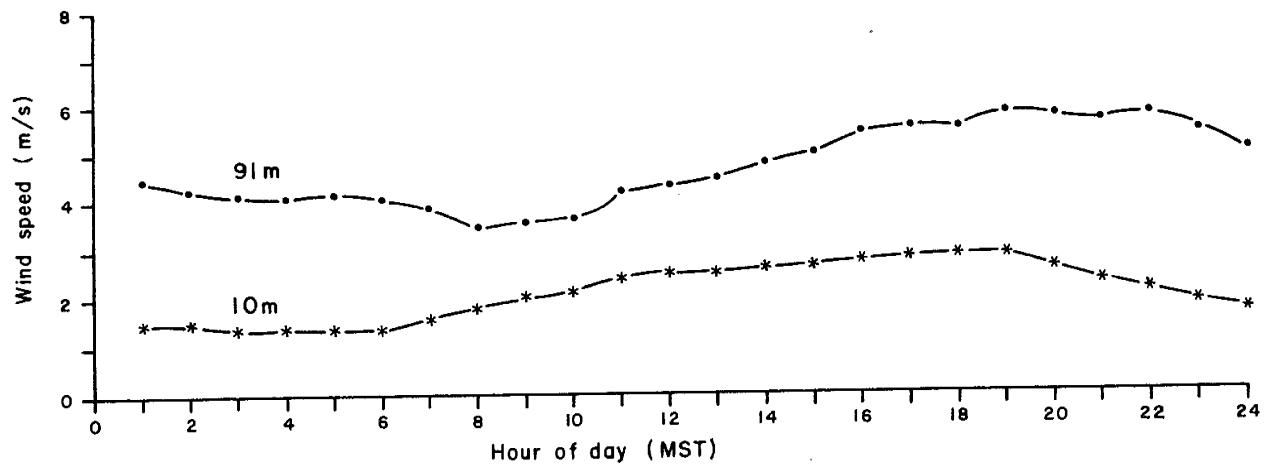
Wind speed is the magnitude of the air velocity vector. *Calm* denotes the absence of measurable wind. Therefore, the occurrence of a calm is highly dependent upon the instrument used to measure the mean wind. "Dead" calms are virtually nonexistent. There is always some gentle motion of air and although this may be immeasurable with instruments, visual observations of smoke or other visible material will reveal the drift.

Wind speeds display a regular diurnal variation with maximums in the afternoon and minimums near sunrise (except in stormy periods). This pattern results from the variations in turbulent mixing which transports momentum downward efficiently in the afternoon, but not in the early morning. Wind speeds at higher elevations show a reversed pattern for the same reason. Figure 1.7 illustrates this behavior.

Wind speeds at a number of U.S. stations are described reasonably well by a log-normal distribution (Luna and Church 1974). When the logarithm of wind speed is



(a) Winter



(b) Summer

Figure 1.7 Diurnal variation of 10-m and 91-m wind speeds in Calgary.

plotted against cumulative percent occurrence on probability paper, a log-normal distribution appears as a straight line. Slight deviations from log-normality were attributed to poor anemometer response at low speeds and to averaging problems at high speeds. The geometric standard deviation derived from the first and third quartiles through the relation

$$\sigma_g = (V_{75}/V_{25})^{0.74129}$$

ranged from 1.4 to 2.5 with a typical value of 1.9. (The geometric standard deviation σ_g is simply the standard deviation of the natural logarithm of the wind speed). A log-normal distribution characterized by the mean speed and a geometric standard deviation of 1.9 may not be universal and caution is advised in applying it to sites of different topography, climate and roughness.

Other workers have used the Weibull distribution (Justus et al. 1976, Stewart and Essenwanger 1978), which gives the probability of a wind speed less than V as

$$P(<V) = 1 - \exp [- (V/c)^k]$$

where $P(<V)$ is the cumulative probability that the wind speed is less than V, c is the scale factor and k is the shape factor. For values of k between 0 and 1 the distribution has its mode at zero and is a decreasing function. For k = 1 the distribution is exponential and for k = 3.5 the distribution is approximately normal (Hennessey 1977).

A three-parameter Weibull distribution may provide a better fit to the data than does a two-parameter Weibull distribution (Stewart and Essenwanger 1978). The third parameter, called the location reference, enters as a constant representing a translation of the origin or zero point on the wind speed scale.

A square-root-normal distribution has been used (Widger 1977), but the fit is not good for the lowest speed ranges. The Weibull distribution is a special case of the generalized gamma distribution which has also been applied (Sherlock 1951). If the shape factor is equal to two, the Weibull distribution reduces to the Rayleigh distribution which has been used by some workers (Hennessey 1977). The bivariate normal (Gaussian) distribution can also be used, but its requirement for five parameters makes it unwieldy. The unimodal two-parameter distributions like the log-normal or Weibull are much easier

to work with and provide a reasonable fit to the data. The Weibull tends to be slightly better than the log-normal (Justus et al. 1976).

Hawrelak et al. (1976) plotted the complementary cumulative distribution for a number of Alberta stations as in Figure 1.8. Maarouf and Fraser (1972) used the Gumbel distribution to estimate extreme winds for 13 Alberta aviation weather stations and calculated return periods as a function of wind direction as in Figure 1.9.

Wind speed persistence refers to the tendency of the wind to remain in the same speed range for consecutive time intervals. This is usually presented as a tabulation of the frequency with which the wind speed was less than certain arbitrary values for various numbers of consecutive hours. Shaw et al. (1972) produced maps of Canada showing the number of occurrences of winds less than 3 m/s lasting 24 to 47 hours and lasting 48 hours and longer (Figures 1.10 and 1.11). This information indicates how often uninterrupted periods of poor volumetric dilution can be expected. Hage and Longley (1970) examined the duration of light wind periods in Calgary and Edmonton (Figure 1.12) and concluded that periods of persistent light winds of a few days duration were not uncommon in winter.

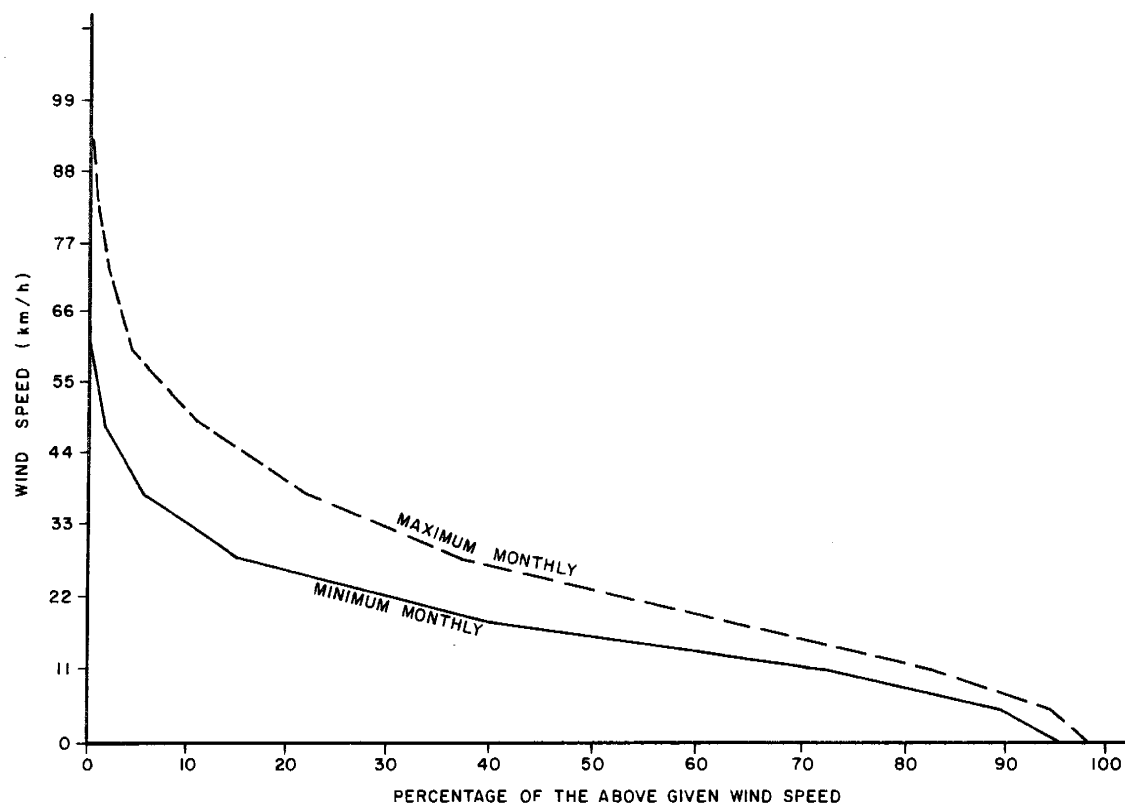
It has been found that the probability of wind speed persistence below fixed reference speeds follows a composite distribution consisting of a power function distribution for the initial portion with a transfer to a scaled exponential distribution (Sigl et al. 1979). The mean run duration in hours (t_{avg}) for 1-h data can be estimated from the ratio of run level (V) to mean wind speed (V_{avg}) by the regression equation

$$\log t_{avg} = 0.009 + 0.803 V/V_{avg}$$

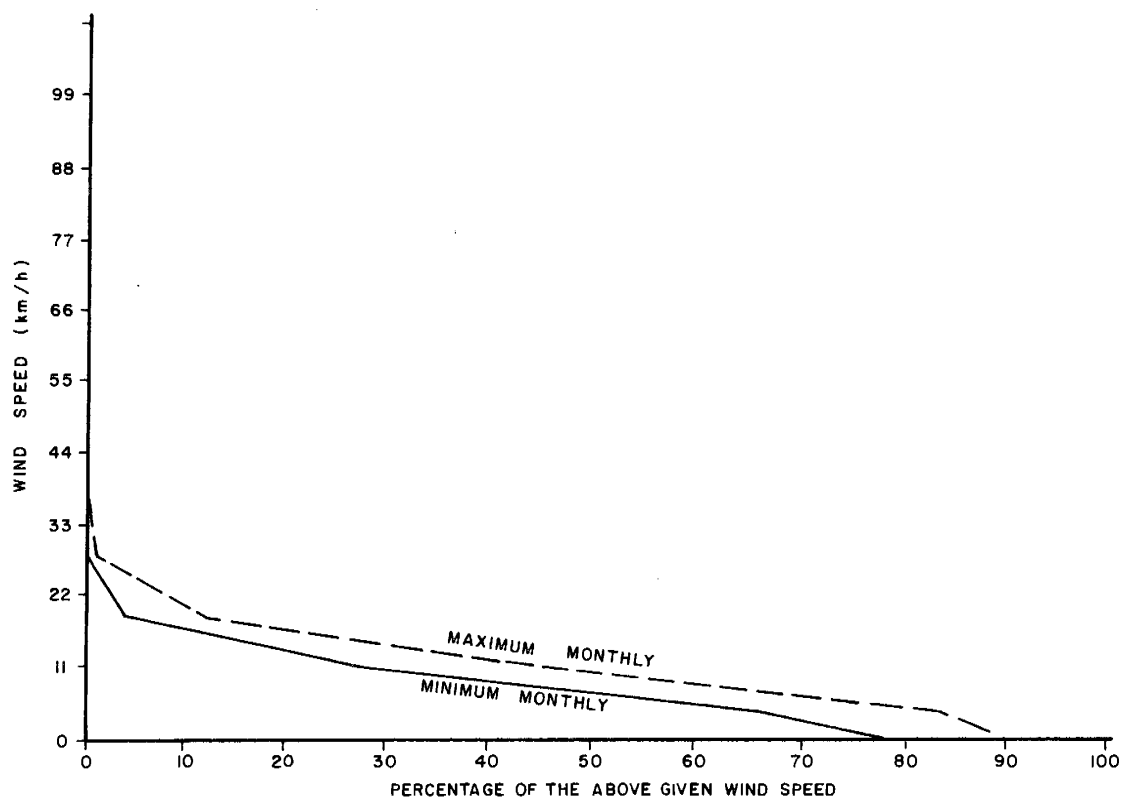
For example, at Edmonton International Airport the mean annual wind speed is 3.9 m/s. If the above relation holds, the average duration of winds below 3 m/s is 4.2 h.

Many of the equations used in air quality meteorology reflect an inverse dependence on wind speed. Average conditions should therefore be defined by using the mean value of the reciprocal of wind speed, $(1/V)_{avg}$, rather than the reciprocal of the mean wind speed, $1/V_{avg}$. Using the log-normal distribution, the ratio between these two quantities for a given wind direction is given by

$$(1/V)_{avg} / (1/V_{avg}) = \exp (\sigma^2 \ln v)$$



(a) Lethbridge



(b) Fort McMurray

Figure 1.8 Percentage of wind speeds above a specified value (after Hawrelak et al. 1976).

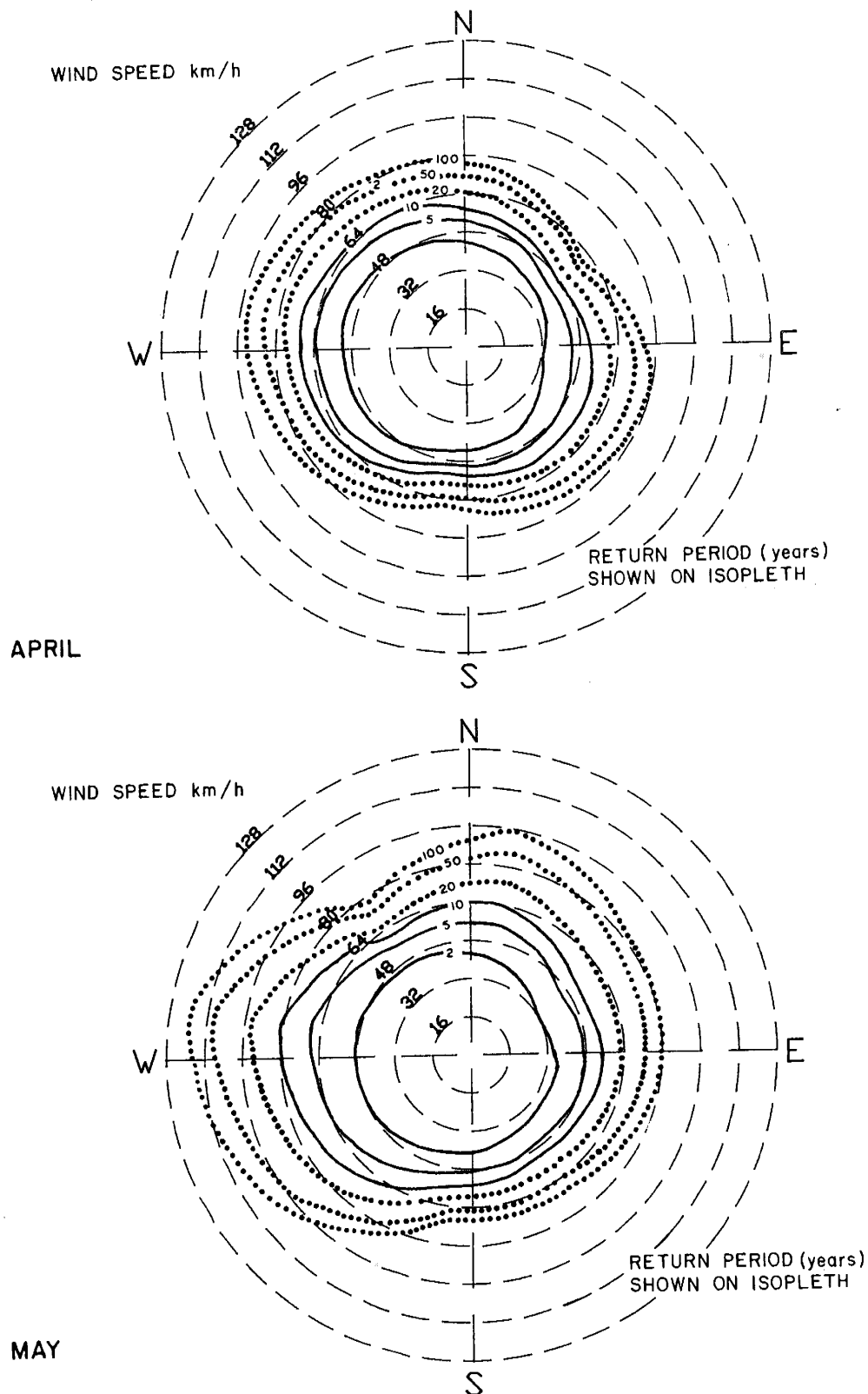
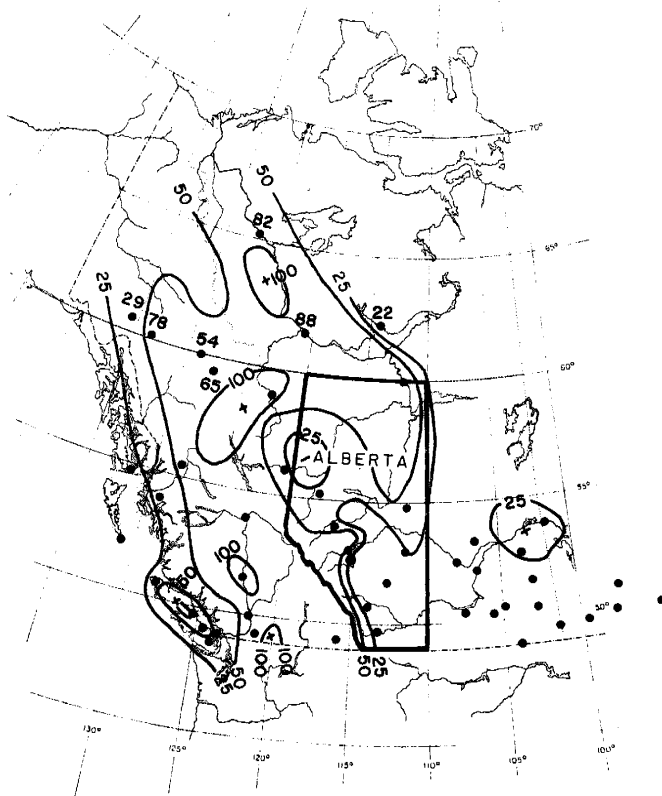
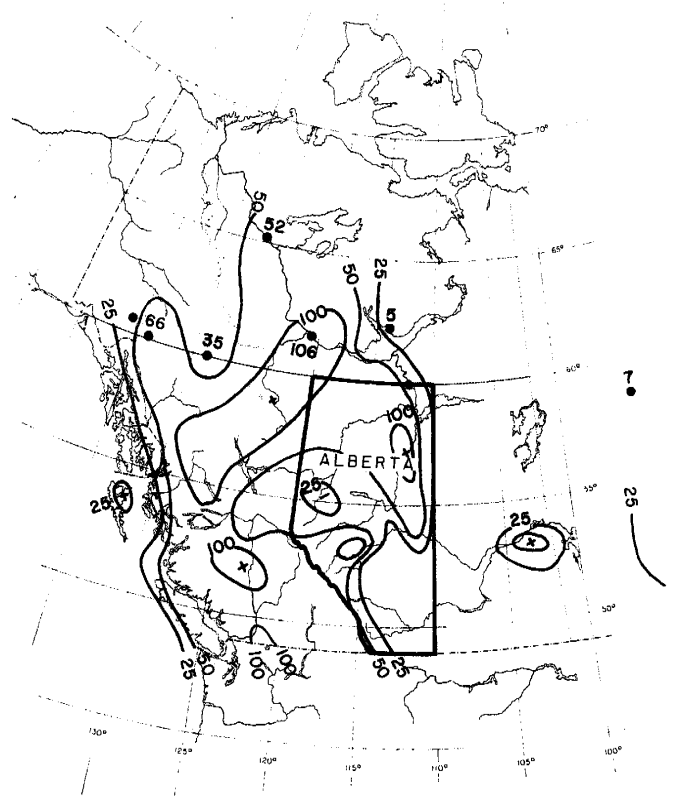


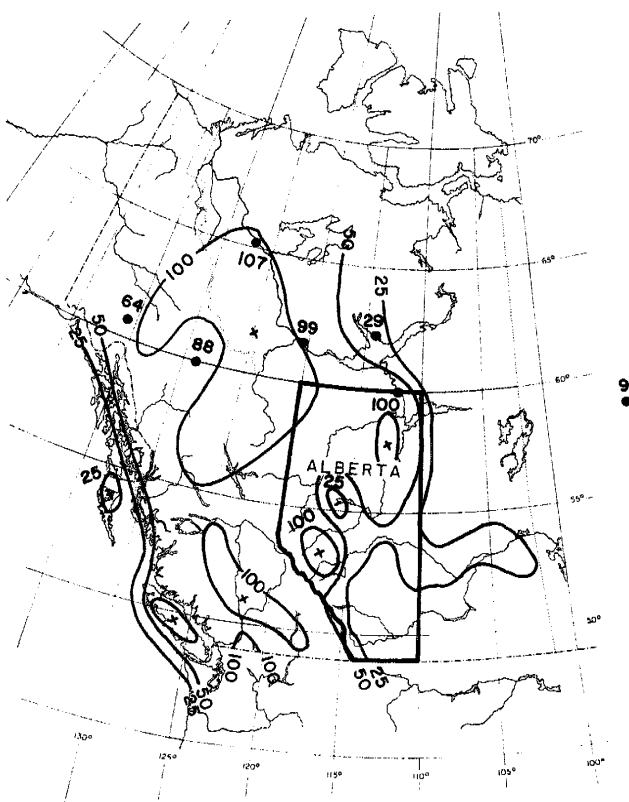
Figure 1.9 Return periods for extreme wind speeds, Medicine Hat 1957-1966 (after Maarouf and Fraser 1972).



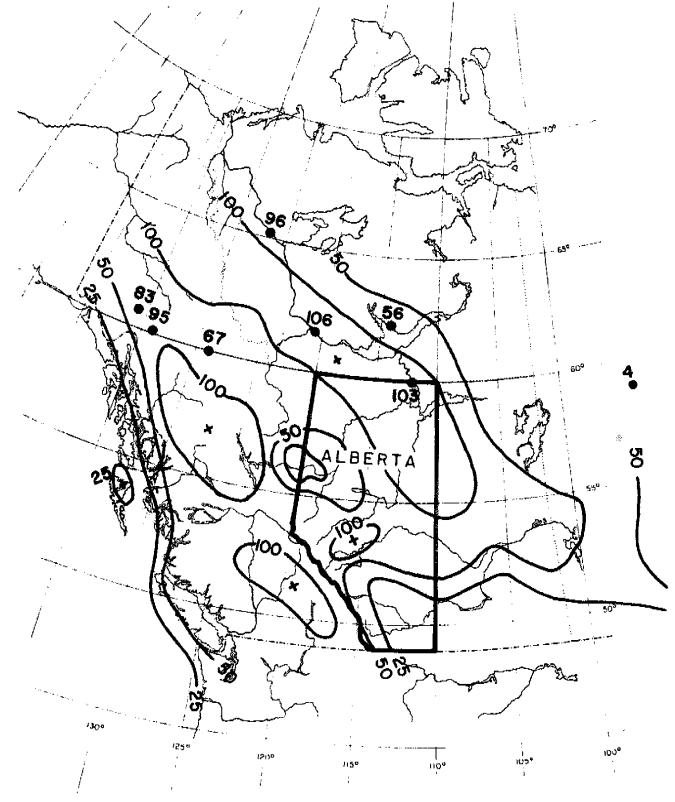
Spring



Summer

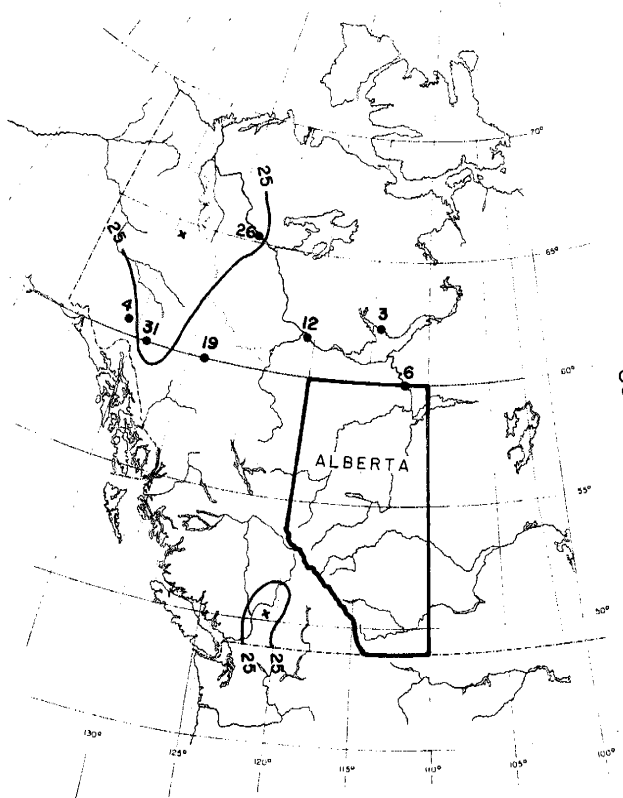


Autumn

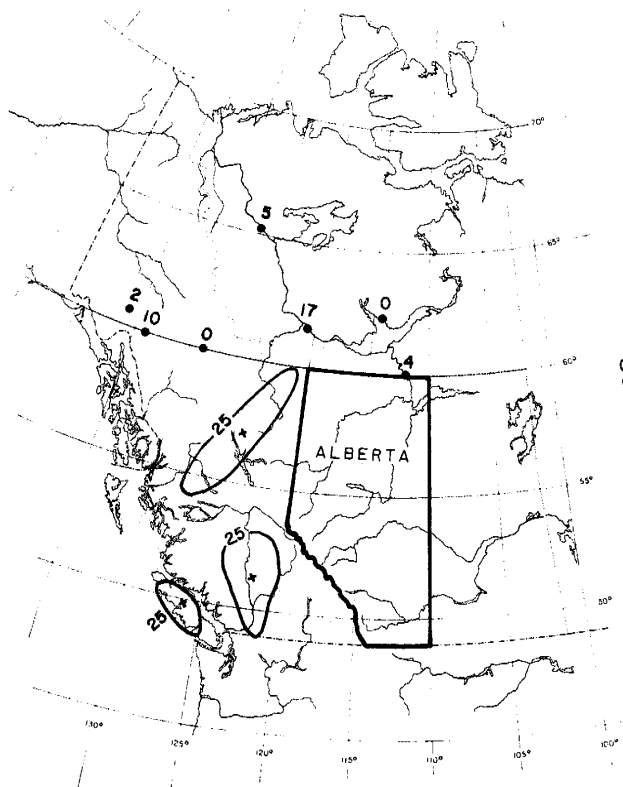


Winter

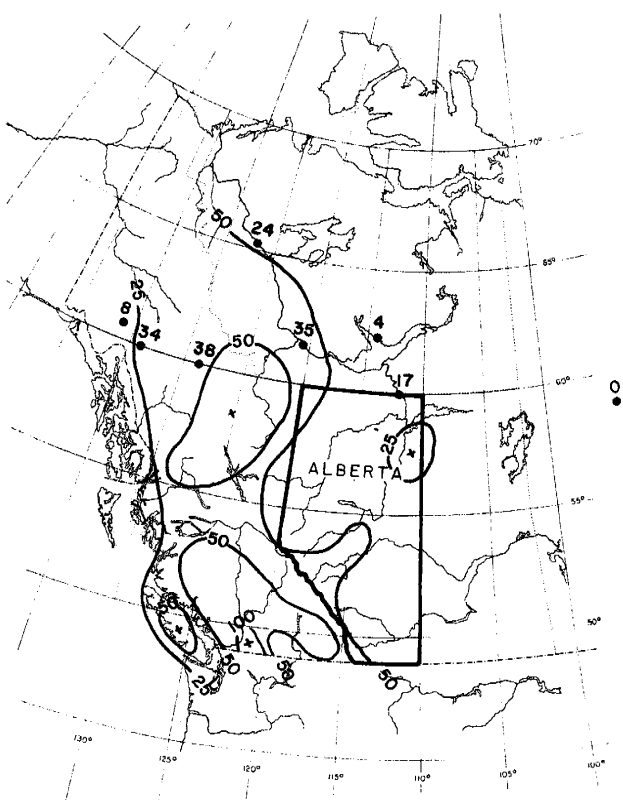
Figure 1.10 Number of occurrences of persistent light winds ($< 3 \text{ m/s}$) lasting 24 to 47 h in years 1957 - 1966 (Shaw et al. 1972).



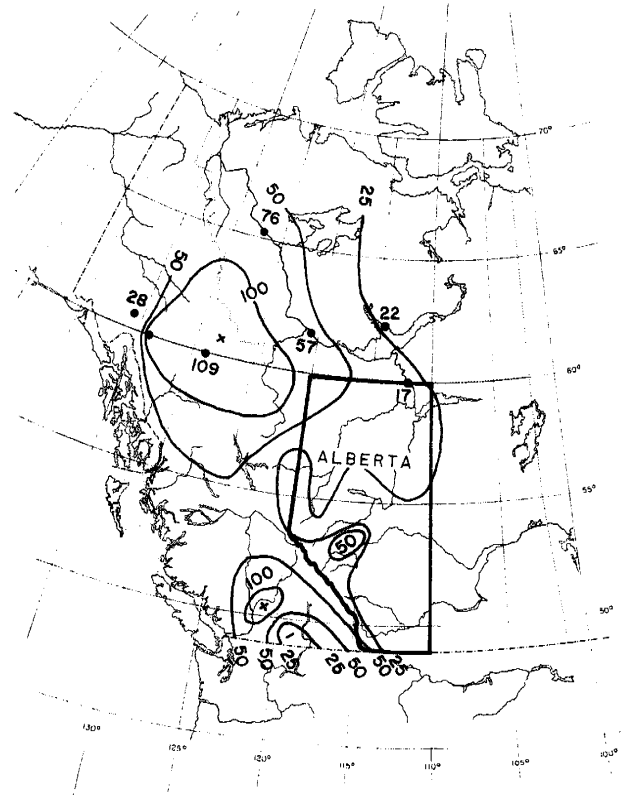
Spring



Summer

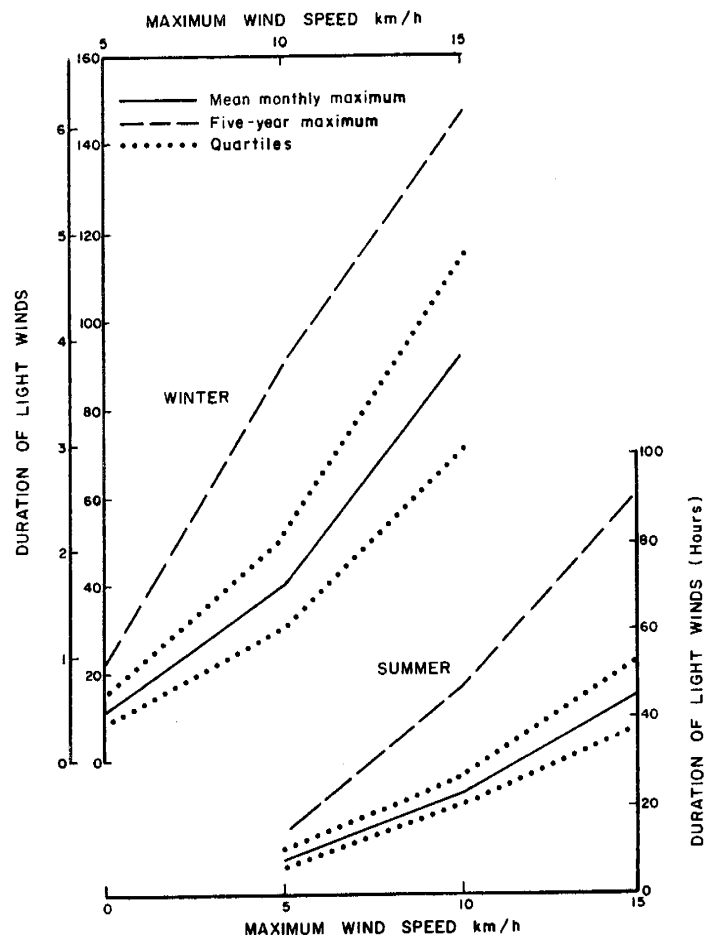


Autumn

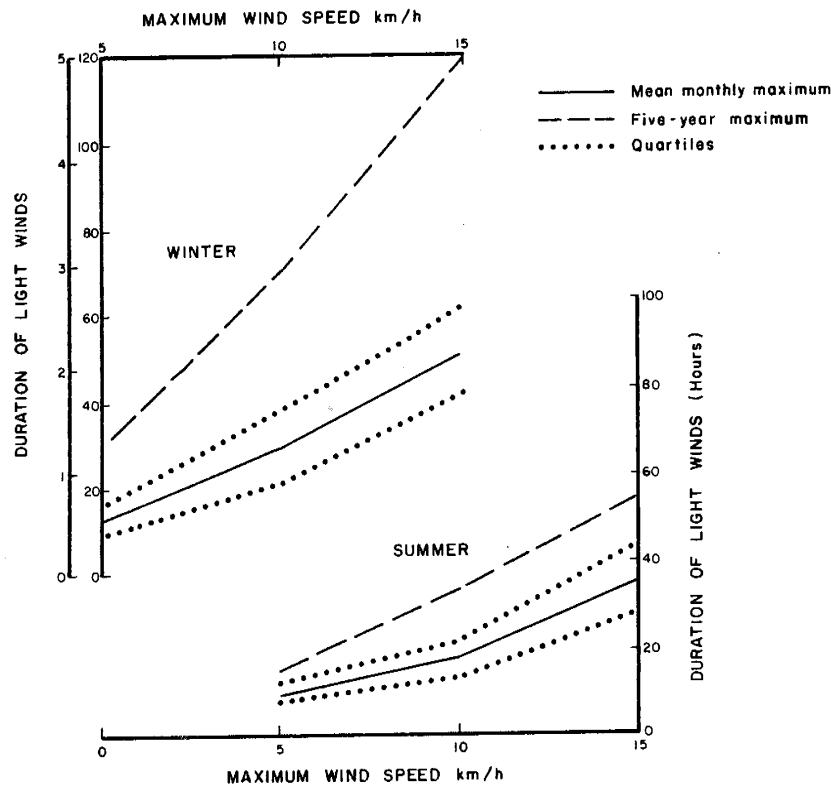


Winter

Figure 1.11 Number of occurrences of persistent light winds (< 3 m/s) lasting 48 h and longer in years 1957-1966 (Shaw et al. 1972).



(a) Edmonton Municipal Airport



(b) Calgary International Airport

Figure 1.12 Duration of light winds in Edmonton and Calgary (Hage and Longley 1970).

where $\sigma^2_{\ln V}$ is the variance of the distribution of $\ln(V)$. A humped gamma distribution applied to the speed of the mean wind or the component of the mean (Neumann 1977) gives wind speed in a specified direction

$$(1/V)_{\text{avg}} = 1/(V_{\text{avg}} [1 - (\sigma/V_{\text{avg}})^2])$$

where σ^2 is the variance of V , the wind speed.

1.1.4 Wind Vector

Using the normal (circular) distribution assumption (Appendix B) Henry (1957) mapped the vector means winds and standard vector deviations of upper winds in Canada. The western portions of his 85-kPa maps, with the units converted to SI, are reproduced in Figures 1.13 and 1.14. The wind component in any direction making an angle θ with the vector mean may be taken as normally distributed with mean value $V_r \cos\theta$ and standard deviation 0.71σ .

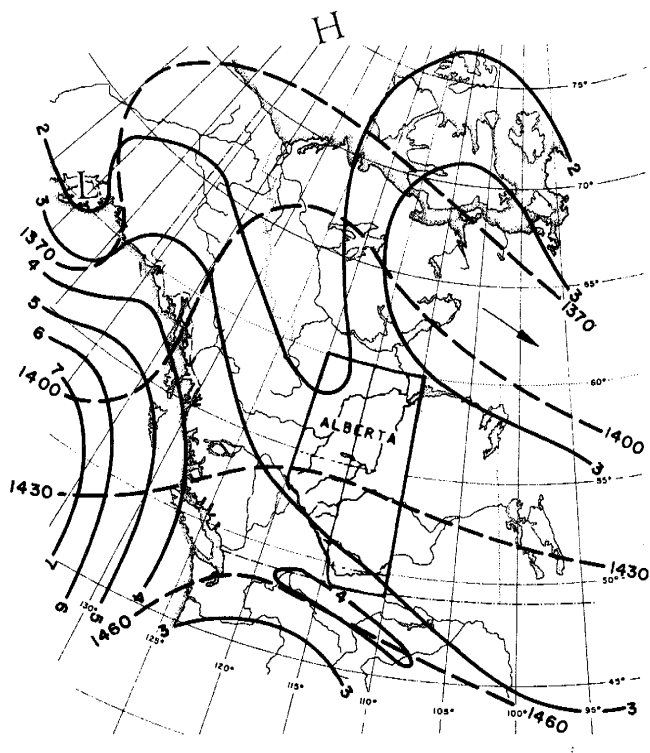
Davenport and Baynes (1972) analyzed Canadian data at altitudes of 300 m and 500 m above ground level for the period 1960 - 69 using the bivariate Gaussian distribution of wind vectors and the Fisher Tippet Type I distribution for extreme annual winds. The resulting patterns of vector mean winds and standard vector deviations were broadly similar to those of Henry for the 85-kPa level (about 1400 m above sea level). The contours of 30-year return wind speeds were similar in pattern but had values about 9 m/s larger than those given in the National Building Code for surface winds.

A measure of the variability of the wind is provided by the *constancy* of the wind is defined as the ratio of the mean vector speed (magnitude of the wind velocity vector) to the mean scalar speed (Singer 1967),

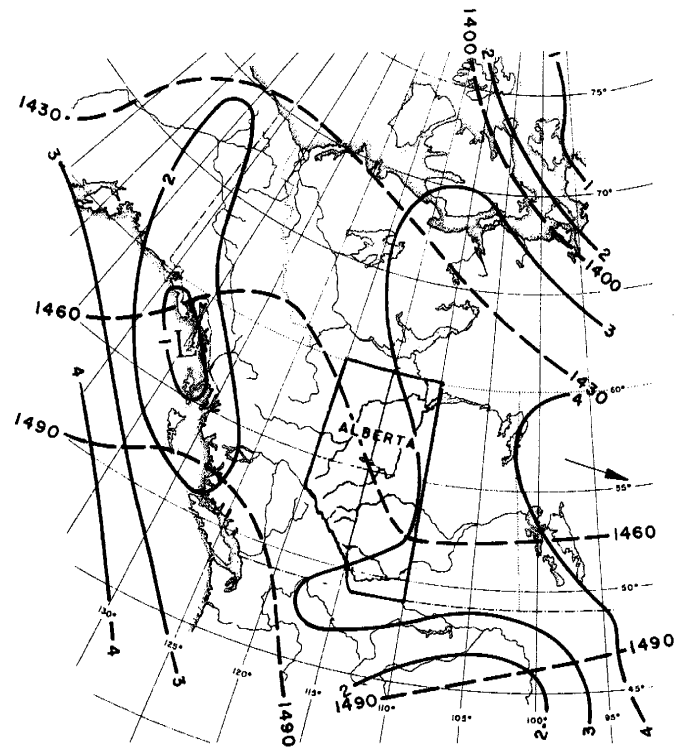
$$k = (V_v)_{\text{avg}}/V_{\text{avg}}$$

A value of one means that the wind direction has not changed over the averaging period and a value of zero suggests a completely symmetrical direction and speed distribution. The constancy decreases in a nonlinear fashion as a function of mean angular range. To linearize this dependence, *steadiness* is defined as

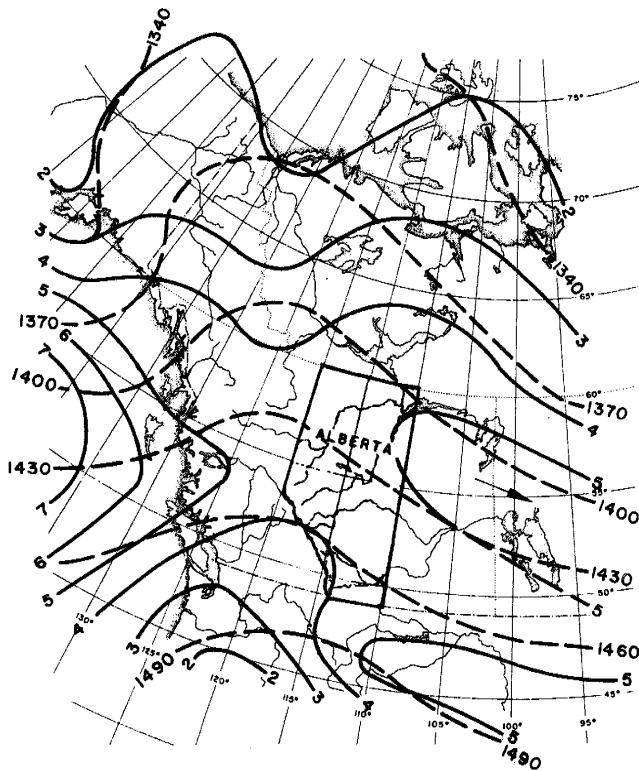
$$S = 2/\pi \sin^{-1} k$$



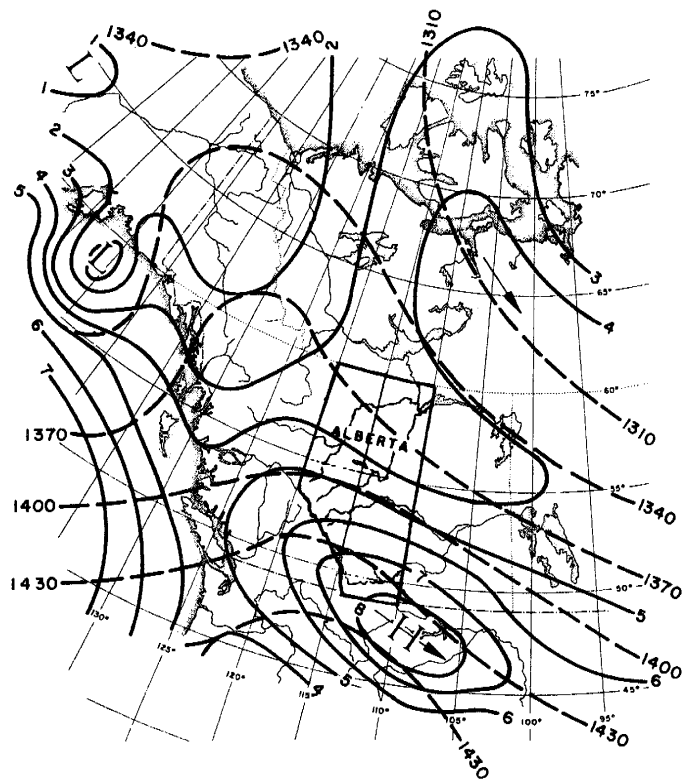
Spring



Summer

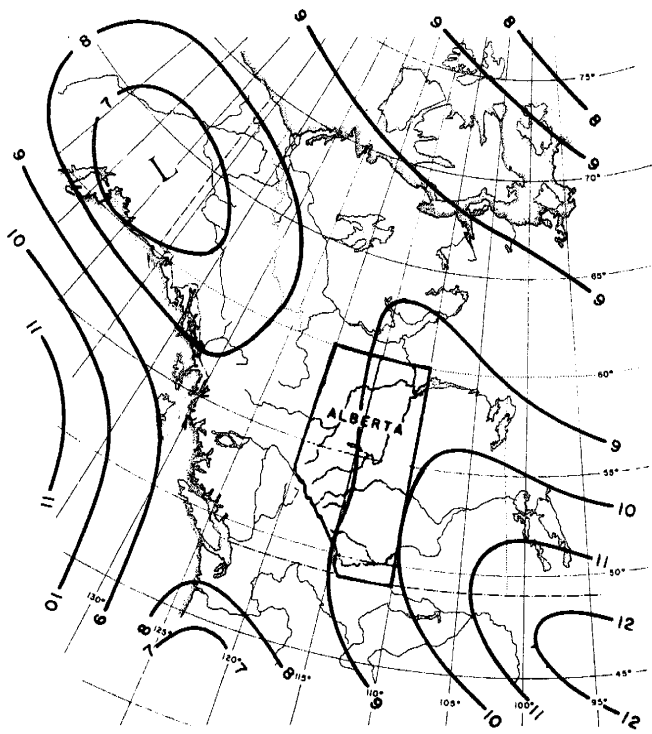


Autumn

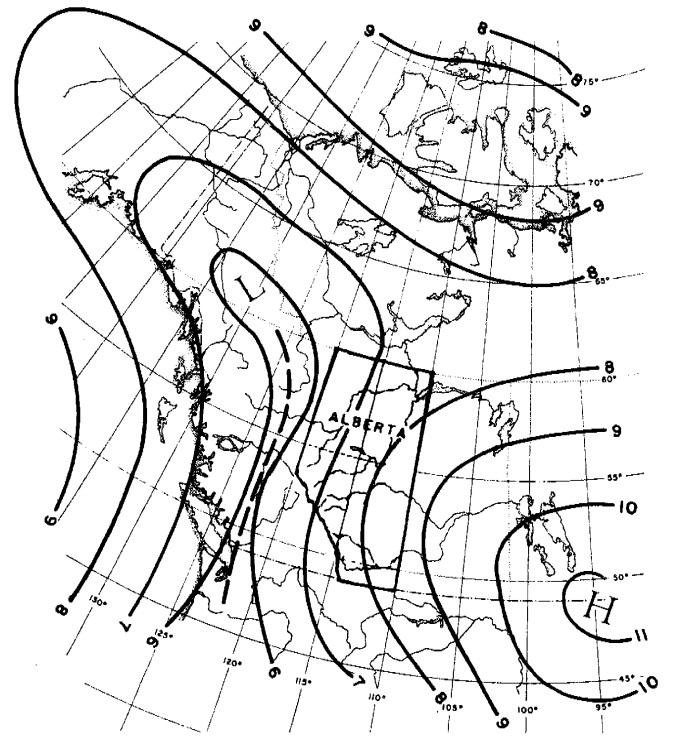


Winter

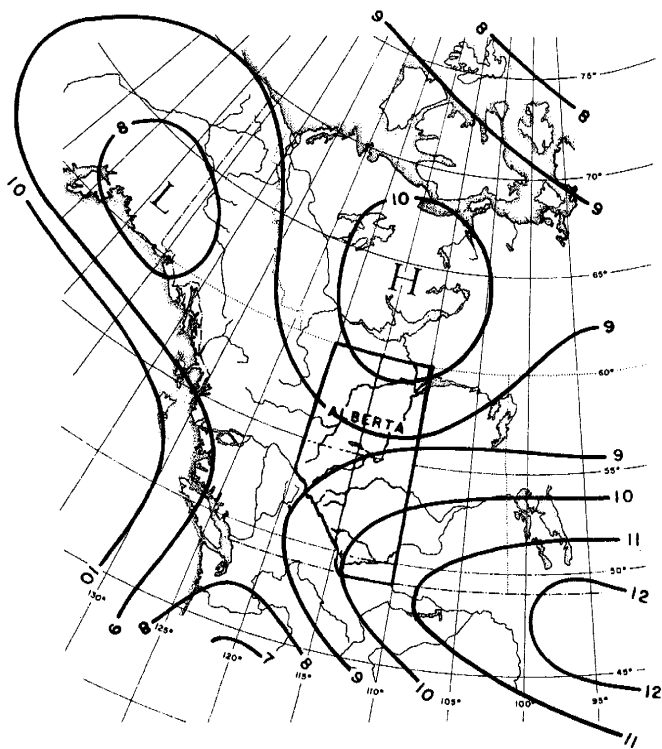
Figure 1.13 Vector mean winds (m/s) at the 85-kPa level for the years 1946-1952 (Henry 1957). Dashed lines are normal geopotential heights in metres.



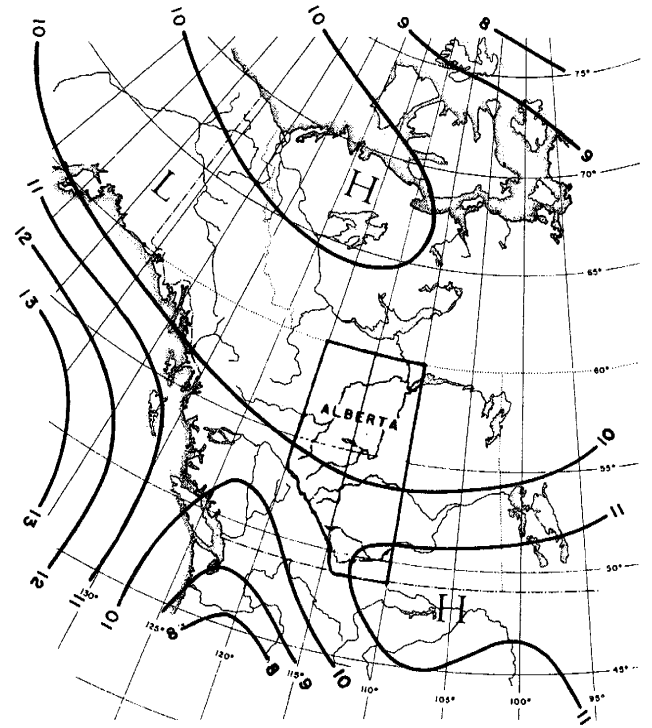
Spring



Summer



Autumn



Winter

Figure 1.14 Standard vector deviation (m/s) of winds at the 85-kPa level for the years 1946 - 1952 (Henry 1957).

where k is the constancy. As a result of this transformation a change of 0.1 in S corresponds to a deviation of 18° . Extreme monthly and annual values of S fit a Fisher Tippet Type I distribution:

$$P(S) = \exp [-e^{-a(S-M)}]$$

where $P(S)$ is the probability of the steadiness being less than S , M is the mode of the distribution and a is the dispersion of the distribution. Thus recurrence intervals of persistent winds may be determined for planning purposes (Singer and Nagle 1970).

If the horizontal wind field is known as a function of time and vertical motion is neglected, the path followed by any individual air parcel can be derived. Such air trajectories can be constructed manually by examining conventional weather maps in succession and assuming that the wind speed and direction affecting the parcel are constant for the period centred on map time. Forward trajectories (Figure 1.15) are constructed to determine where an air parcel from a given region will arrive after a specified time has elapsed. Backward trajectories (Figure 1.16) are constructed to determine the origin of an air parcel that has arrived at a given location. Denison (1977) used surface weather maps drawn every 6 h and 85-kPa weather maps drawn every 12 h to prepare a climatology of low-level air trajectories for the Alberta oil sands area. The seasonal results for non-precipitation days are reproduced in Figure 1.17 and 1.18. Peake et al. (1985) used trajectories to explain the observations of peroxyacetyl nitrate at a number of stations in the Edmonton area. Based on trajectories, Legge (1988) attributed high ozone concentrations on Fortress Mountain to transport from British Columbia and the state of Washington.

1.2 TEMPERATURE STRATIFICATION

Organized vertical motion in the lower atmosphere over flat terrain is caused by buoyancy forces whose magnitude and direction are determined by the difference in density between a given air parcel and the surrounding air. The variation of density with height will control the behavior of the parcel over a period of time as it changes its altitude in response to local buoyancy forces. If the environment of the air parcel is in equilibrium, the density is inversely proportional to the temperature (according to the ideal gas law). Density profiles may therefore be represented by temperature profiles.

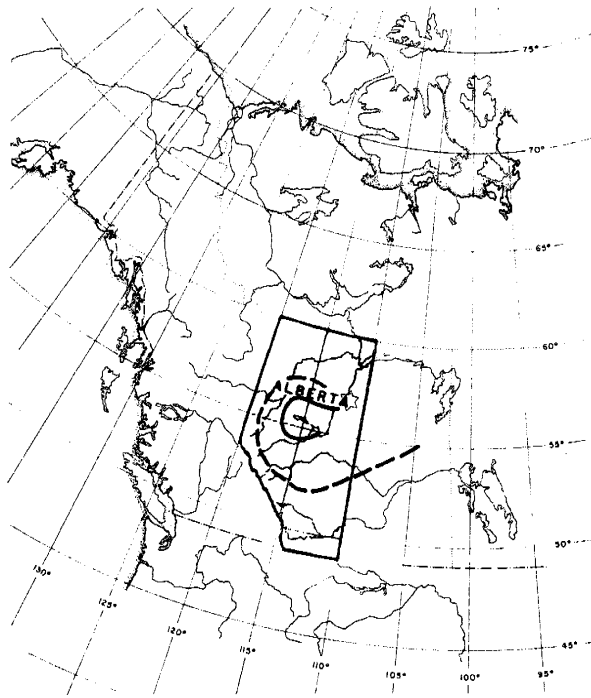


Figure 1.15 Trajectories of air parcels leaving the Alberta oil sands area for the 72-h period ending March 11, 1973, 00:00Z (after Denison 1977). Solid line is surface; dashed line is 85-kPa.

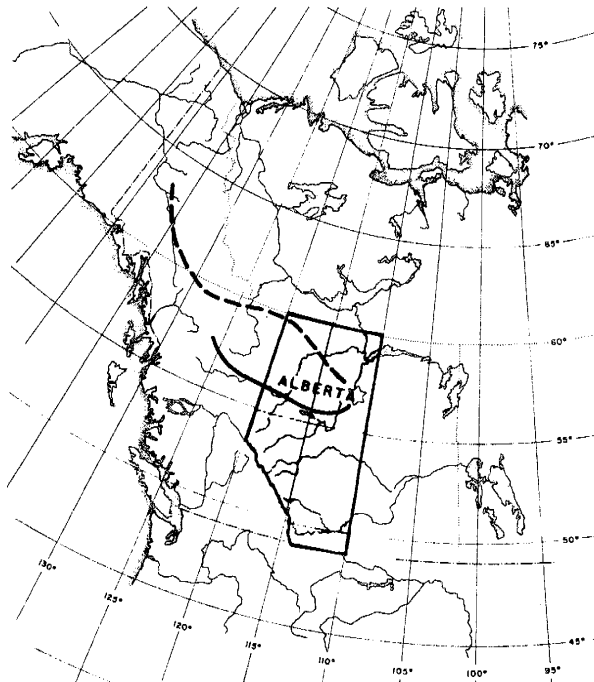
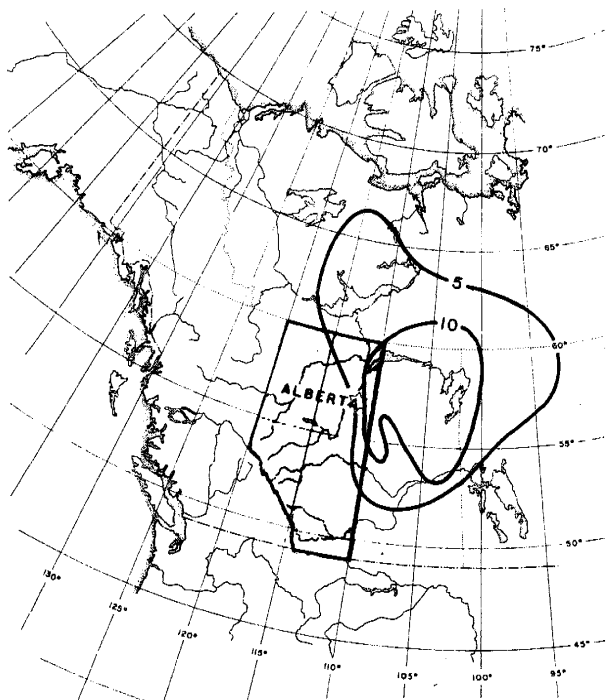
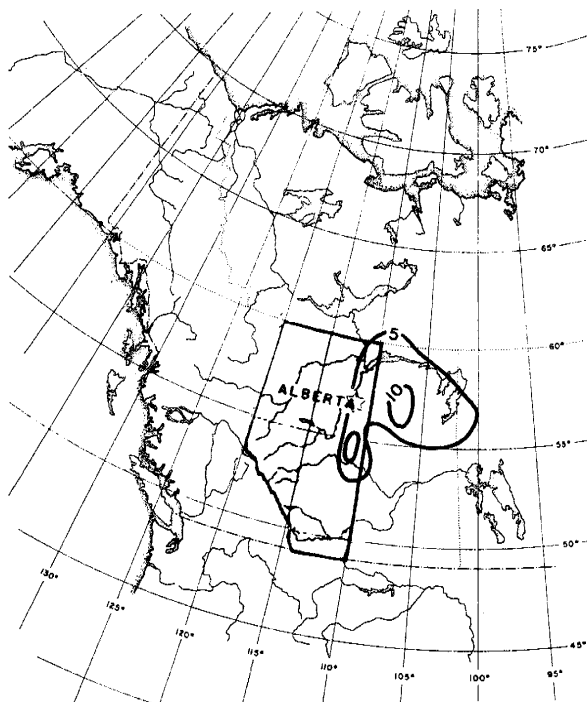


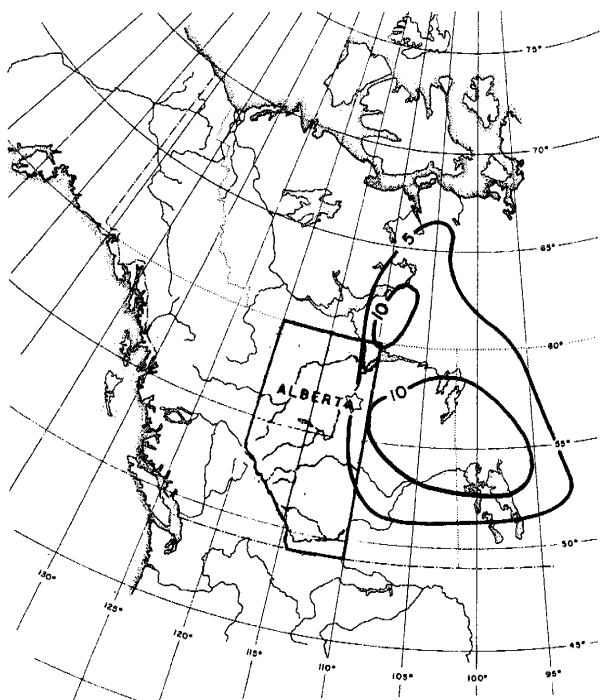
Figure 1.16 Trajectories of air parcels entering the Alberta oil sands area for the 72-h period ending February 2, 1975, 00:00Z (after Denison 1977). Solid line is surface; dashed line is 85-kPa.



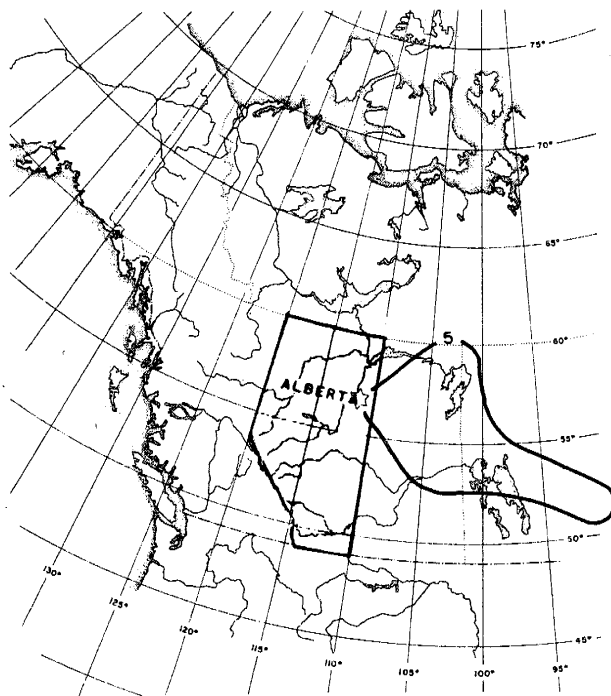
Spring



Summer

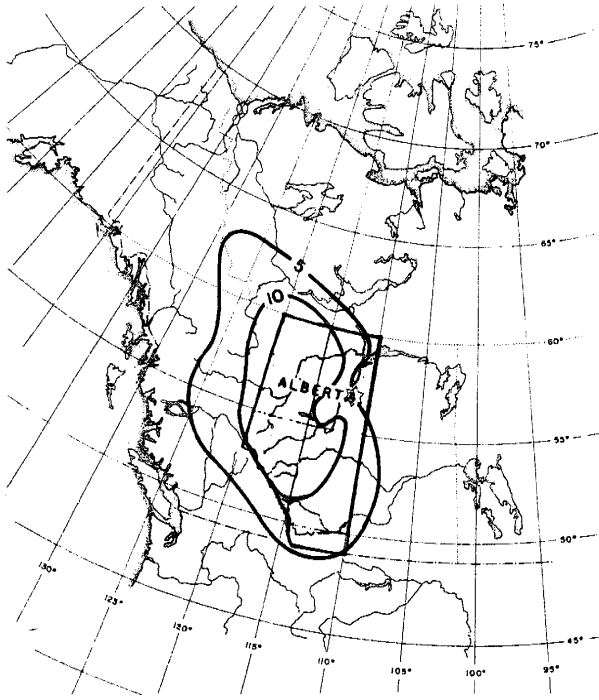


Autumn

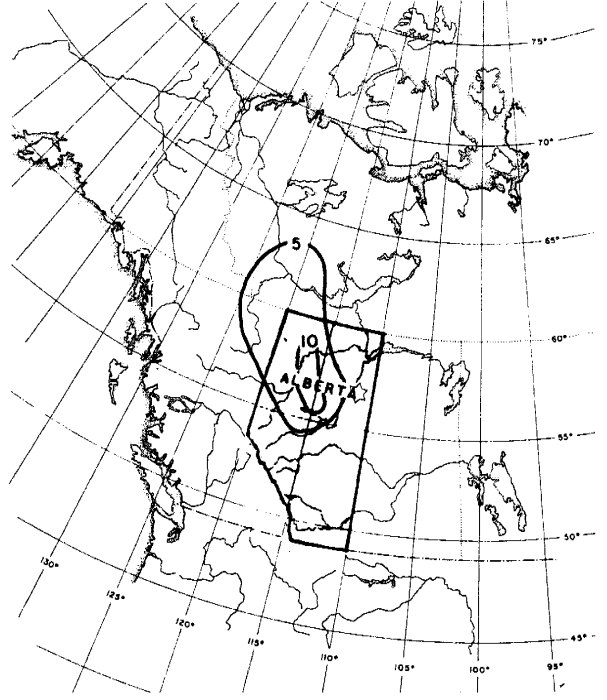


Winter

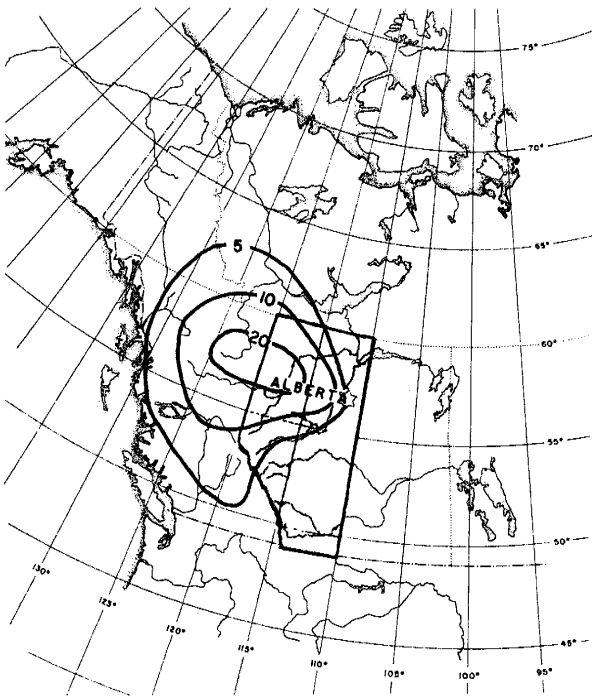
Figure 1.17 Frequencies of 72-h air parcel motions out of the Alberta oil sands area at the 85-kPa level (after Denison 1977).



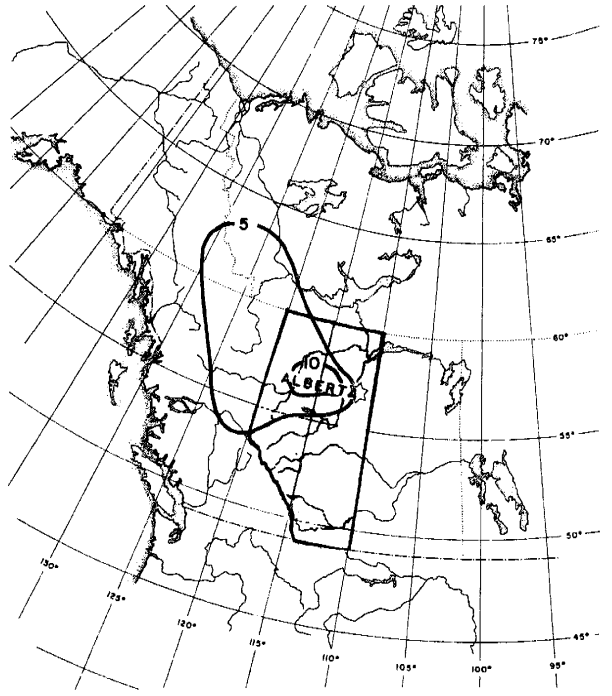
Spring



Summer



Autumn



Winter

Figure 1.18 Frequencies of 72-h air parcel motions into the Alberta oil sands area at the 85-kPa level (after Denison 1977).

The rate of change of temperature with height is known as the *temperature gradient*. Often the negative of the temperature gradient, called the *lapse rate*, is used to talk about the rate of decrease of temperature with height.

1.2.1 Hydrostatic Stability

When a parcel of air is forced upwards in the atmosphere it will encounter lower pressure, expand and cool. Assuming (a) there is no heat exchange with the environment, (b) the parcel maintains its identity, (c) there are no compensating motions in the environment and (d) there is no phase change of the water vapor in the parcel, the rate of cooling can be calculated to be close to $0.01^{\circ}\text{C}/\text{m}$. This is known as the *dry adiabatic lapse rate* and it constitutes a convenient reference for the classification of atmospheric temperature stratification.

When an air layer has a dry adiabatic lapse rate, displaced parcels will always have the same temperature (density) as their environment and hence will tend to remain at whatever level they find themselves. Such a condition is said to be neutral because vertical motion is neither encouraged nor discouraged.

When the temperature in an air layer decreases more rapidly than $0.01^{\circ}\text{C}/\text{m}$ the layer has a *superadiabatic lapse rate*. A parcel displaced upwards will be warmer (less dense) than its environment. Buoyancy forces will tend to accelerate the upward motion. Similarly, a parcel displaced downward will be cooler (more dense) than its environment and (negative) buoyancy forces will tend to accelerate the downward motion. Since vertical motion in both directions is encouraged, such a condition is said to be *unstable*. Small changes in air parcel temperature can result in large vertical movements.

When the temperature in an air layer decreases more slowly than $0.01^{\circ}\text{C}/\text{m}$, the layer has a *subadiabatic lapse rate*. A parcel of air displaced upward will be cooler (more dense) than the environment. The buoyancy force will oppose this motion and tend to return the parcel to its original level. Similarly a downward displaced parcel will be warmer (less dense) than the environment and buoyancy will act to restore the parcel back to its original height. Because vertical motion is

inhibited, such a condition is said to be *stable*. Small changes in air parcel temperature will produce only small movements to equilibrium heights where parcel and environment temperatures are equal. Two special situations can be identified:

- (1) Where the lapse rate is zero, an air layer is said to be *isothermal*
- (2) Where the lapse rate is negative and temperature increases with height, an *inversion layer* is said to be present. Vertical motion is strongly suppressed.

The potential temperature gradient is often a more convenient index of hydrostatic stability. *Potential temperature* is defined as the temperature which a parcel of air would have if brought adiabatically from its original pressure in the atmosphere to sea level (actually 100 kPa). The potential temperature gradient equals the temperature gradient plus the dry adiabatic lapse rate. A hydrostatically neutral layer will have a potential temperature gradient of zero; a hydrostatically unstable layer, a negative value; and a hydrostatically stable layer, a positive value.

If an air parcel is moist, upward displacement and expansion may cool it below the dewpoint. Condensation will then occur and latent heat will be released. Thus the temperature decrease will be less than for dry air. The *moist adiabatic lapse rate*, also called the *wet adiabatic lapse rate*, the *saturated adiabatic lapse rate* and the *saturated pseudo-adiabatic lapse rate* is not a constant, but is a function of temperature and pressure. In the lowest kilometre of the atmosphere the moist adiabatic lapse rate ranges from about 0.009°C/m at -30°C to about 0.004°C/m at +20°C. The moisture consideration leads to a somewhat more complicated classification of hydrostatic stability. If γ is the environmental lapse rate, γ_d is the dry adiabatic lapse rate and γ_s is the moist adiabatic lapse rate, then an air layer is said to be:

absolutely stable for $\gamma < \gamma_s$
 saturated-neutral for $\gamma = \gamma_s$
 conditionally unstable for $\gamma_s < \gamma < \gamma_d$
 dry-neutral for $\gamma = \gamma_d$
 absolutely unstable for $\gamma > \gamma_d$

If an entire layer of air moves in the vertical, the stability of the layer with respect to parcel displacements will be modified. If the layer is unsaturated both before and after the altitude change, ascent will cause the lapse rate to approach dry adiabatic while descent will make the lapse rate depart more from dry adiabatic. In other words, lifting a dry layer destabilizes it while sinking stabilizes it with respect to parcel displacements.

If an initially unsaturated layer is lifted until it becomes saturated its stability depends upon the gradient of equivalent potential temperature. *Equivalent potential temperature* is defined as the potential temperature that an air parcel would have if all the water vapor were condensed out and the latent heat used to warm the air. If this gradient is equal to zero, the layer is *convectively neutral*; if greater than zero, *convectively stable*; if less than zero, *convectively unstable*. Sometimes convective stability is referred to as potential stability.

In the conditionally unstable case, the layer is stable with respect to upward movements of saturated (moist) air parcels. If an initially unsaturated air parcel is forced upward it will first cool at the dry adiabatic rate until it reaches the *lifting condensation level*. It will then cool at the moist adiabatic rate until at some point its temperature will equal that of the environment. Above this *level of free convection* the parcel will experience a positive buoyant force that will accelerate it upward.

The foregoing discussion has referred to hydrostatic stability as determined by the *parcel method*, that is, by considering the state of equilibrium of a sample air parcel displaced a short distance from its initial position. If the displaced parcel is subject to restoring forces, the layer is termed stable, if it is subject to forces which accelerate it, unstable; if subject to no net force, neutral. Hydrostatic stability, also called *static stability*, *thermal stability*, *vertical stability*, and *convectational stability*, describes a state of the atmosphere. It does not express any particular motion of an air parcel, nor does it imply existence or the nature of any forces that may cause parcel displacements; it only suggests the probable effect on an air parcel.

The parcel method does not allow for compensating motions to take place in the environment of a moving parcel, an assumption which is invalid on continuity grounds. Only if the ascending parcel is very small and isolated will the descending motion of the environment be small enough to be neglected. Compensating motions can be accounted for in a procedure called the *slice method*.

The concept of hydrostatic stability is based on two key assumptions: (1) hydrostatic equilibrium exists, in other words, there is a balance between all vertical forces and there is no acceleration of air in the vertical, and (2) individual air parcels maintain their identity and do not mix with their surroundings. The first is a reasonable approximation since vertical accelerations are usually small compared with gravity. The second must be qualified because observations indicate that entrainment occurs.

Use of the term "stability", by itself with no adjectives, has caused considerable confusion in the field of plume dispersion meteorology. When a weather forecaster speaks of "stability" he usually means "convective stability", especially as it relates to cumulus cloud development. When speaking of turbulence, the word "stability" has quite a different meaning (section 1.3.4).

1.2.2 Formation of Elevated Inversions

A number of different situations can produce inversion layers not connected to the earth's surface:

- (1) Differential advection (horizontal motion) - Overrunning by warm air and undercutting by cold air, singly or together produces a statically stable layer. The *frontal inversion* that occurs at the interface between two airmasses is of this type. The *chinook inversion* forms when warm Pacific air crosses the mountains and overlays cold Arctic air near the ground.
- (2) Differential vertical motion - If a statically stable air layer descends and the descent is accompanied by a horizontal outflow, then a *subsidence inversion* forms. The inversion then becomes strongest not where the largest descent occurs, but where the vertical shrinking is the greatest.

The most significant subsidence inversions are associated with the semipermanent anticyclones located near 30°N and 30°S latitude. The air between about 500 and 5000 m descends at a rate of about 1000 m per day in these subtropical anticyclones, giving inversion bases that can vary from 300 m to a few thousand metres above ground, with a thickness ranging from 500 to 1000 m.

Ascending motion combined with lateral divergence and/or sinking of the environment can also lead to stabilization. If an air parcel rises through the condensation level it cools at the moist adiabatic lapse rate whereas the compensating downward motion in the environment is accompanied by heating at the dry adiabatic rate. The result is the formation of a *cloud-base inversion*.

The sinking of air in advance of a front can lead to a small inversion below the frontal inversion, which for obvious reasons is called a *pre-frontal subsidence inversion* (Sweeney 1979).

- (3) Differential radiation - Clouds radiate energy as blackbodies. If the air above them is relatively dry, upward radiation escapes and the cloud top cools while the air above remains unaffected. Therefore the layer above the cloud is stabilized and possibly becomes an inversion. A similar situation occurs when a dry stratum lies above a moist one or above a haze layer. In order to obtain such a moisture discontinuity in the first place a stable layer would likely have to be present. Thus differential radiational cooling is generally most effective in strengthening or maintaining existing inversions.
- (4) Differential mixing - If a statically stable layer is subjected to strong mixing, the upper portion cools and the lower portion warms. This may make the upper portion of the mixed layer cooler than the layer immediately above or the lower portion warmer than the layer immediately below. In either case a *turbulence inversion* is formed.

- (5) Differential evaporation and condensation - As warm moist air rises in rain clouds, condensation occurs releasing latent heat which adds to the warmth and to the upward motion. As cool dry air sinks to replace the warm air, some rain may evaporate into it, causing additional cooling. The resulting temperature differences can be large.

1.2.3 Formation of Ground-Based Inversions

Inversion layers extending upward from the surface are created by two main mechanisms:

- (1) Radiation cooling - At night the surface of the earth radiates energy. If the sky is clear a large amount of heat is lost to space. The air in immediate contact with the surface is also cooled and mixing extends this cooling upward, creating the *nocturnal inversion*. However, if the mixing is too strong the downward heat transfer prevents inversion formation. Light winds, cloudless skies and long nights favor the development of deep, strong radiation inversions. Extreme cooling over the snow-covered regions of northern Canada during long polar nights produces the *Arctic inversion* which can extend to a height of 1000 m.
- (2) Contact or advective cooling - If warm air moves over a cold surface as with the approach and passage of a warm front, heat is removed from the air in contact with the surface. Mixing extends the cooling upward and leads to the formation of an *advection inversion*. Cold water bodies or melting snow can act as such heat sinks. Evaporation of rainwater from the ground is also effective in producing a cold surface. If a warm airstream is retarded near the ground and progressively warmer air is brought in aloft, an inversion can also be formed, even though there has been no loss of heat from the air.

1.2.4 The Significance of Inversions

Although "inversion" has come to be synonymous with pollution problems, this is a misleading generalization. In reality the base height, thickness and strength of the

inversion relative to the source of pollutants must be considered. For example, a ground-based inversion a few hundred metres thick certainly creates a potential air pollution situation in an urban area where most sources of pollutants are at or near ground level. However, the plumes from even moderately tall stacks are likely to rise above the inversion and disperse in the less stable air above. The ground-based inversion will then effectively block any industrial pollution from reaching the ground. With thick ground-based inversions, the plumes from tall stacks may become embedded. Strong inversions with large potential temperature gradients greatly reduce plume rise. Because there is little vertical mixing the pollutants diffuse downward slowly, eventually reaching ground level at some large distance. This situation also sets up the potential for a "fumigation" should the inversion layer be broken up quickly as by strong solar heating (section 2.7).

Elevated inversions are responsible for reducing the dispersion of plumes from tall stacks. If the base of the inversion lies slightly above the level of the plume, then the volume of air available for dilution is severely limited. The elevated inversion acts as a lid restricting mixing in the vertical, reducing dilution and increasing ground-level concentrations.

An elevated inversion is characterized by a base height, a thickness and a potential temperature difference through the layer. These parameters can be combined in various ways to represent the influence of the inversion on air pollution by means of a single numerical value. An index for the California subsidence inversion has been defined (Stanford Research Institute 1958) as

$$I = (\Delta\theta/\Delta Z) \Delta\theta/Z$$

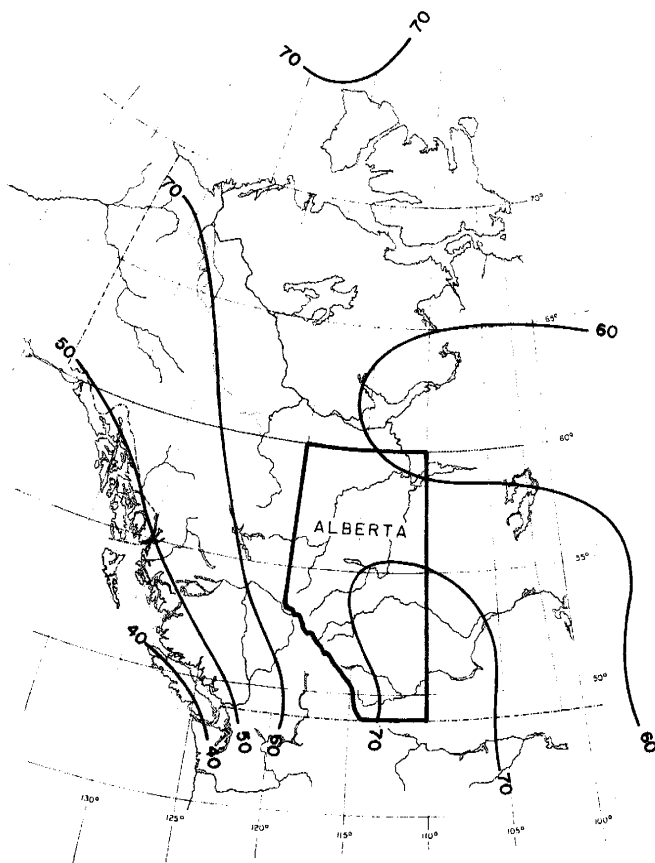
$$= \frac{(\Delta\theta)^2}{Z\Delta Z}$$

where I is an index representing the "intensity" of the inversion, $\Delta\theta$ is the potential temperature difference across the inversion of thickness ΔZ and Z is the height of the inversion base. Large values of I indicated conditions most effective in suppressing the upward dispersion of pollution. In practice a constant was added to the denominator to

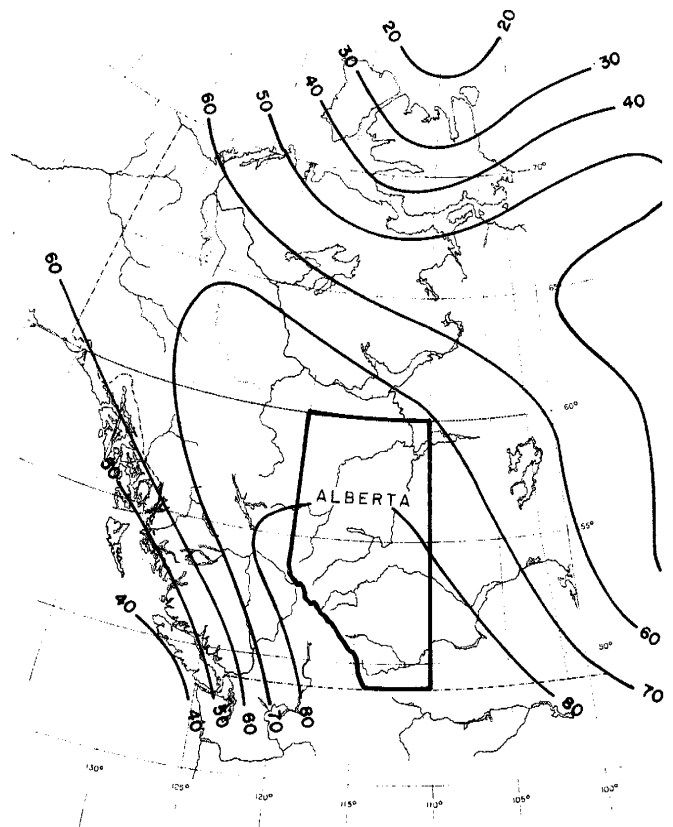
allow for "leakage" through the inversion. Such composite variables permit simple tabulations of frequency and persistence. If the index is significantly correlated with pollutant concentrations, then it is also useful in calculating or predicting concentrations.

Munn et al. (1970) drew maps of ground-based inversion frequencies from Canadian upper air data as shown in Figures 1.19 and 1.20. At 05:00 MST in Alberta, ground-based inversions are present more than 80% of the time in summer and 60% of the time in winter. At 17:00 MST the frequencies are less than 5% in spring and summer and about 40% in winter. Munn and Emslie (1964) showed the frequency and intensity of early-morning ground-based inversions at the old Edmonton Industrial upper air station (Figure 1.21). Median inversion strength was $3^{\circ}\text{C}/1000\text{ ft.}$ or $0.01^{\circ}\text{C}/\text{m}$ in winter and 95% of inversions were less than $13^{\circ}\text{C}/1000\text{ ft.}$ or $0.04^{\circ}\text{C}/\text{m}$. It is more difficult to analyze elevated inversions because the temperature profile is often complex, comprising multiple inversion layers of varying thicknesses and separations. Arbitrary but objective simplifications are required (Holzworth 1974) in order to summarize inversion frequencies by thickness, base height and strength. Kociuba, et al. (1977) developed an inversion typing scheme for the Edmonton (Stony Plain) soundings (Figure 1.22). The frequencies of ground-based and elevated inversions with bases below 85 kPa are summarized in Table 1.1. Examination of the associated weather patterns revealed that the vast majority of elevated inversions with bases below 85 kPa were frontal in origin. Ground-based inversions were present 70% of the time at 05:00 MST and 15% of the time at 17:00 MST. Elevated inversions were present 13% of the time at 05:00 MST and 18% of the time at 17:00 MST. Ground-based inversions were typically 240 m deep with a gradient of $0.073^{\circ}\text{C}/\text{m}$. Elevated inversion were typically 480 m above the surface and 200 m thick with a gradient of $0.018^{\circ}\text{C}/\text{m}$.

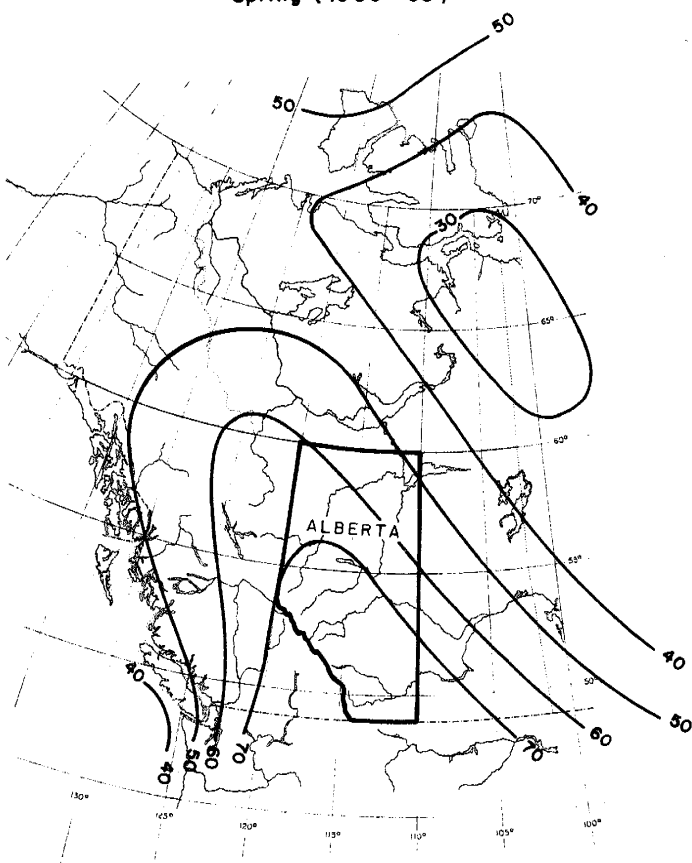
Inversion persistence, despite its obvious importance, has not received much attention because of classification problems and the long interval (12 h) between radiosonde ascents. Analysis methods based on occurrences within fixed altitude bands led to fragmentation of a persistence episode when the inversion oscillated across the boundaries (Sweeney 1979). Different selection criteria can better follow the path of an inversion through various atmospheric levels; inversions rise and fall, coalesce and divide, disappear and reappear, often without obvious cause.



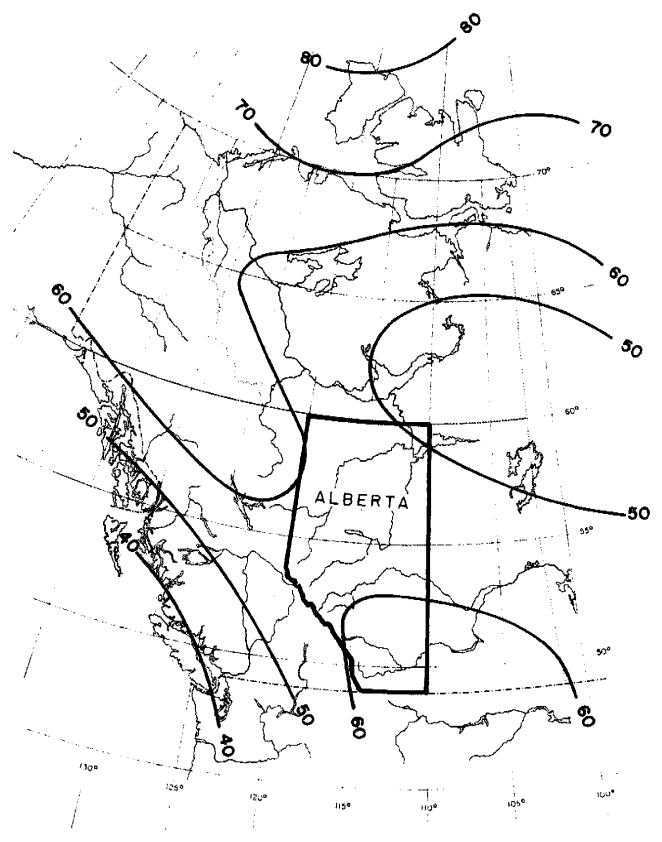
Spring (1966 - 68)



Summer (July , 1965 - June , 1968)

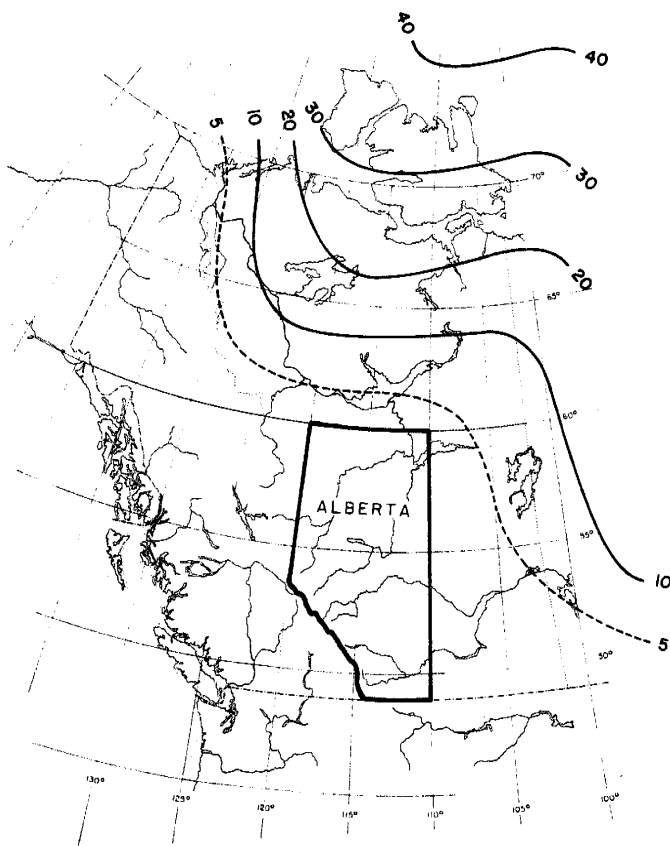


Autumn (1965 - 67)

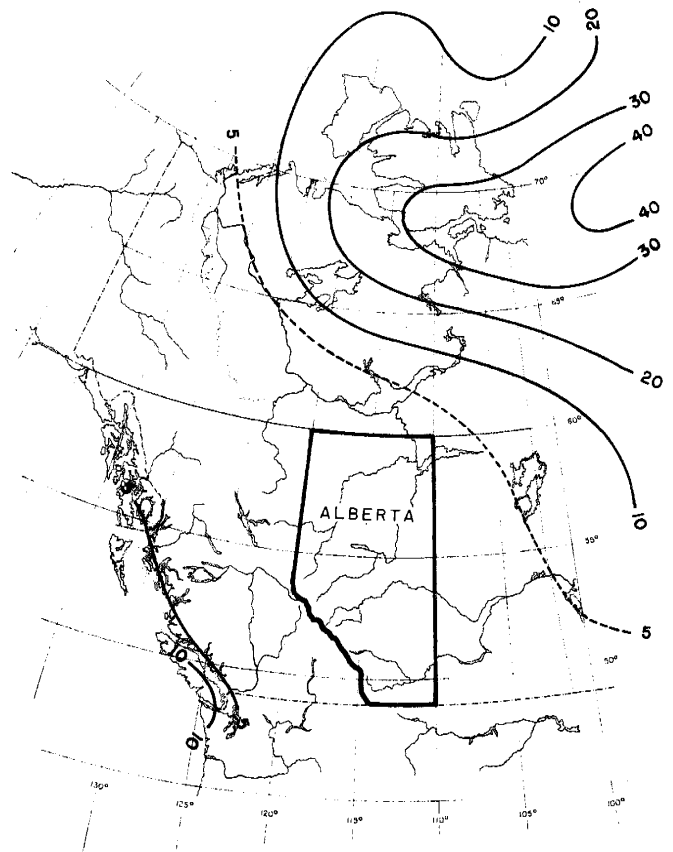


Winter (1966 - 68)

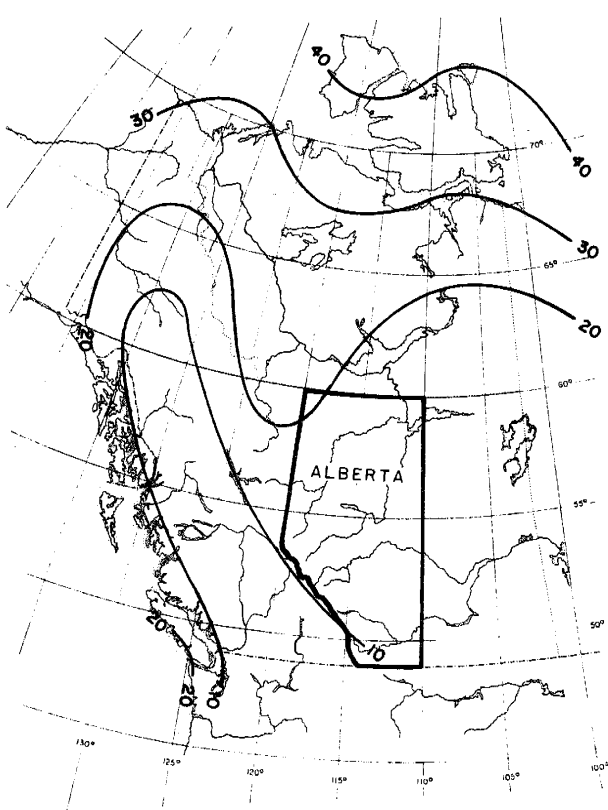
Figure 1.19 Percent frequencies of ground-based inversions at 05:00 MST (Munn et al. 1970).



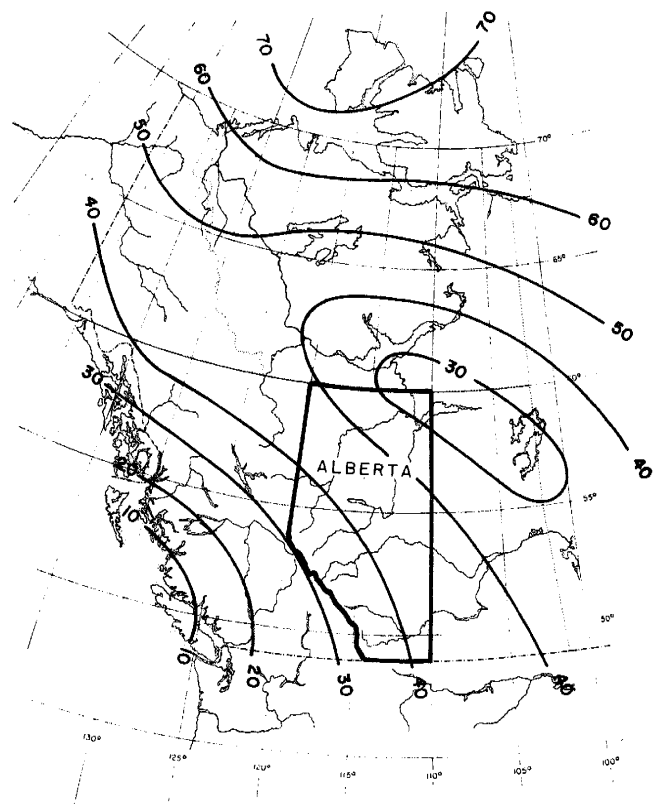
Spring (1966 - 68)



Summer (July, 1965 - June, 1968)



Autumn (1965 - 67)



Winter (1966 - 68)

Figure 1.20 Percent frequencies of ground-based inversions at 17:00 MST. (Munn et al. 1970).

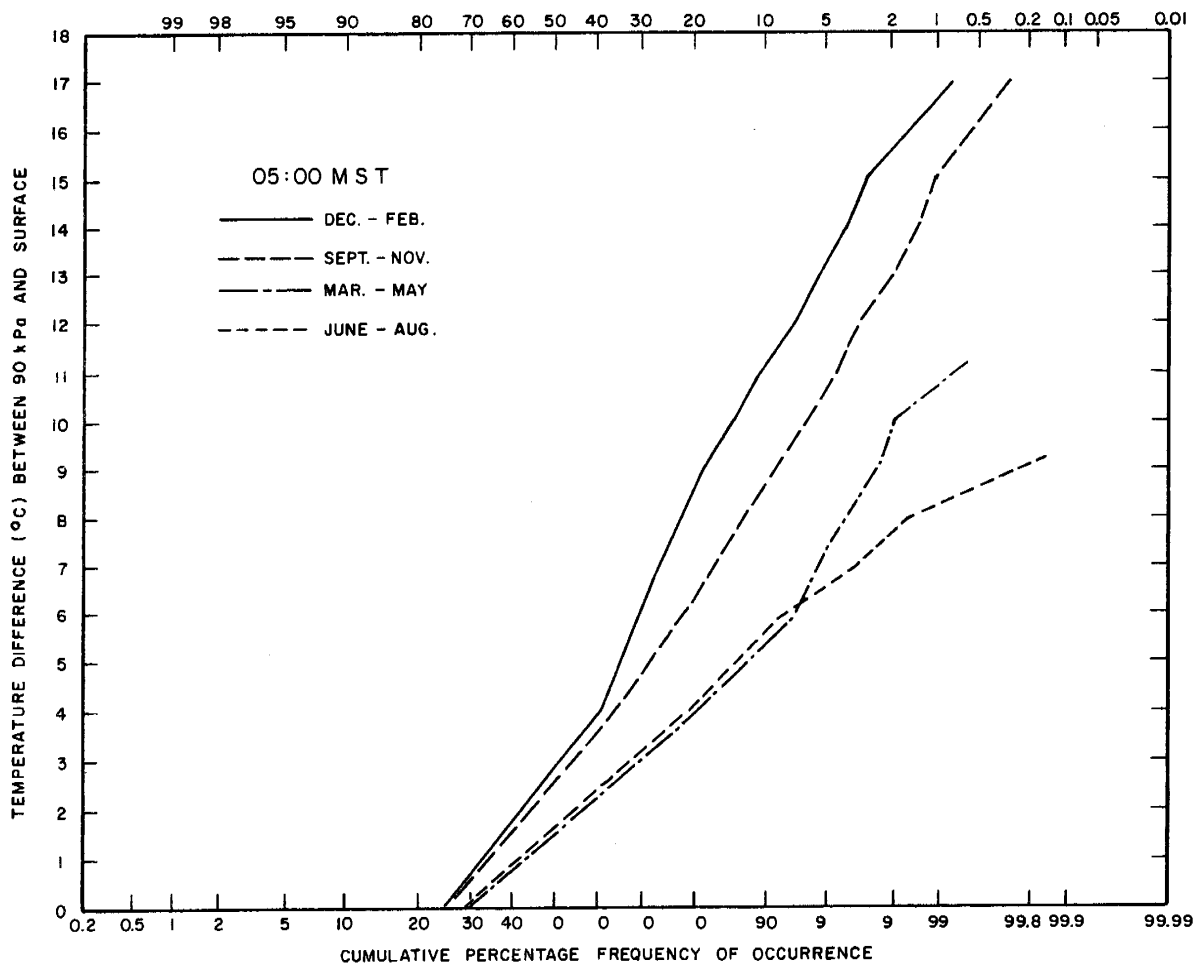


Figure 1.21 Seasonal frequency of ground-based inversions at 05:00 MST Edmonton, June 1957 to November 1960 (Munn and Emslie 1964).

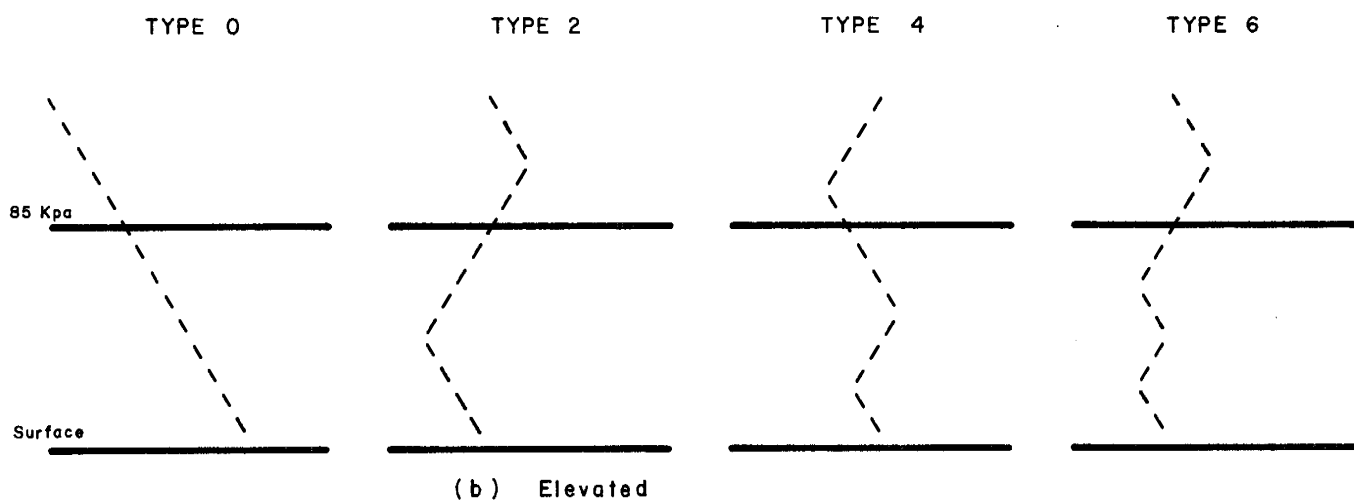
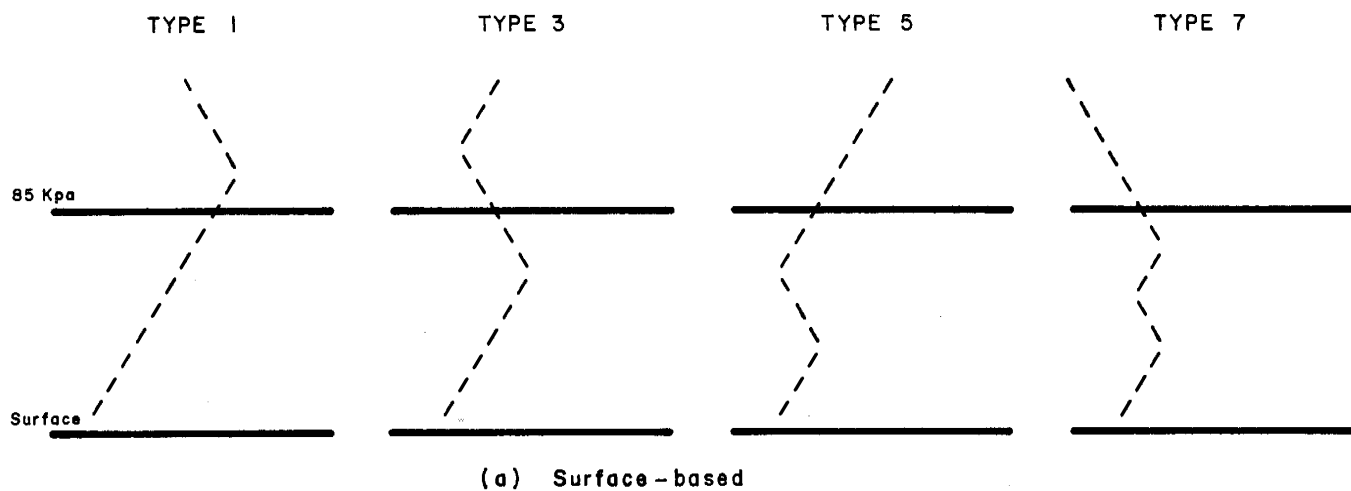


Figure 1.22 Inversion types (Kociuba et al. 1977).

Table 1.1 Averages of inversion types for Edmonton 1971 - 1975
(Kociuba et al. 1977).

Type No.	%	average gradient (°C/100 m)	standard deviation (°C/100 m)	average base above ground (m)	standard deviation (m)	average top above ground (m)	standard deviation (m)	average thickness (m)	standard deviation (m)
17:00 MST									
0	66.3	.3	.1	--	--	5598.9	3791.7	--	--
1	1.3	.6	.4	--	--	928.8	364.9	928.8	364.9
2	10.0	1.2	1.2	401.2	153.1	953.8	273.7	552.6	284.0
3	13.2	1.7	1.6	--	--	257.6	131.6	257.6	131.6
4	7.9	1.5	1.6	264.0	129.1	479.1	128.2	215.1	104.4
5	.5	1.3	1.1	--	--	232.9	139.8	232.9	139.8
6	.3	1.8	1.3	228.8	93.1	402.4	104.8	173.7	113.0
7	.4	1.1	.9	--	--	183.0	112.1	183.0	112.1
05:00 MST									
0	17.3	.7	.1	--	--	4569.5	3697.7	--	--
1	2.9	.8	.7	--	--	851.4	293.9	851.4	293.9
2	7.1	1.1	1.2	357.9	159.0	964.7	310.3	606.8	352.1
3	62.2	2.3	1.8	--	--	241.1	126.0	241.1	126.0
4	6.0	1.6	1.6	252.6	118.4	483.9	117.3	231.3	112.6
5	2.8	2.4	2.0	--	--	182.1	94.1	182.1	94.1
6	.3	1.9	1.3	187.8	69.5	360.2	62.9	172.5	94.2
7	1.3	2.2	1.8	--	--	156.8	60.7	156.8	60.7

Cumulative frequency distributions of temperature gradient for the Edmonton CBC and Calgary Bonnybrook towers are given in Figure 1.23, from Choukalos (1985). In Calgary 95% of inversions had gradients less than $0.08^{\circ}\text{C}/\text{m}$ while in Edmonton 95% of gradients were less than $0.07^{\circ}\text{C}/\text{m}$. The joint distributions of temperature gradient with wind direction are given in Tables 1.2 and 1.3. Strong inversions are most frequent with winds from the S, SW or W in Edmonton whereas in Calgary there is no strong directional preference and the highest frequency is associated with "calms". Average diurnal variations are depicted in Figure 1.24.

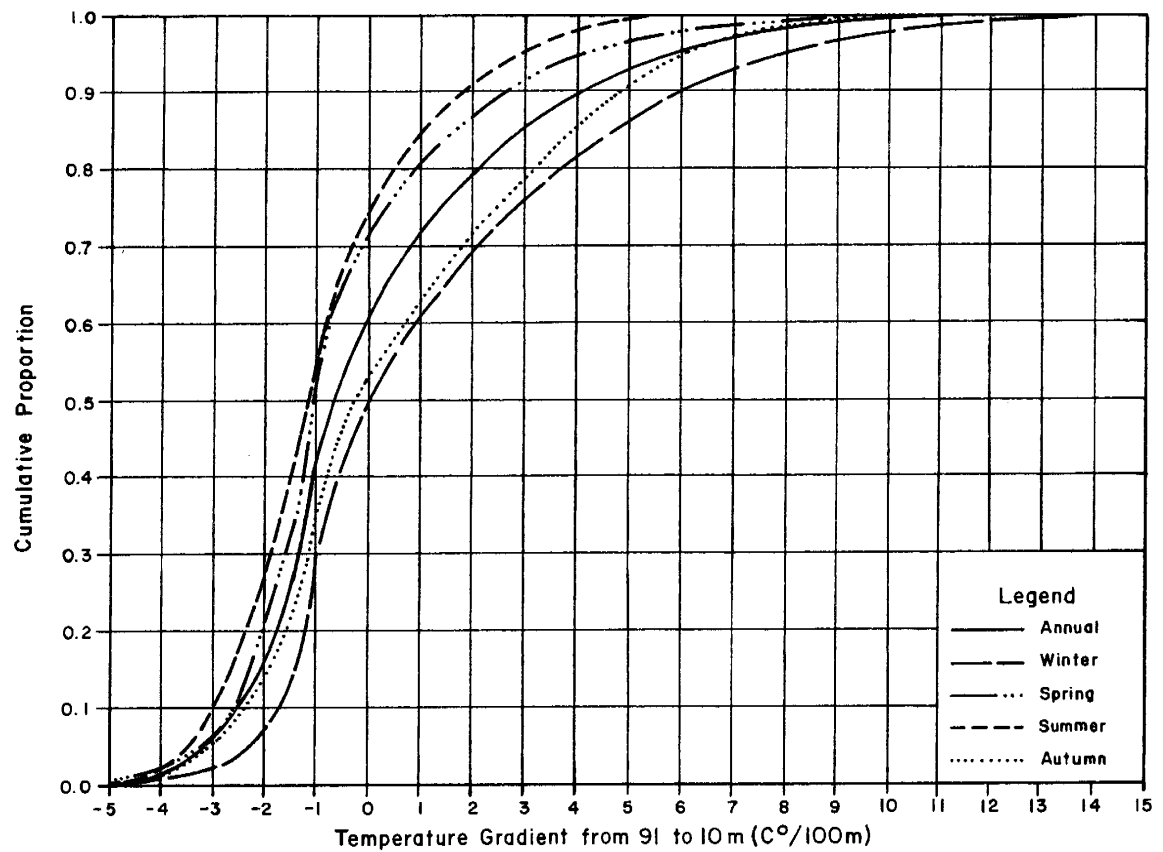
The average temperature gradient measured over 7 years southwest of Edmonton as reported by Sakiyama and Wong (1987) is shown in Table 1.4. The strongest positive gradients occurred near midnight in December with an hourly value of $0.081^{\circ}\text{C}/\text{m}$. The duration of temperature inversions was 20 h in December and only 9 h in June.

The frequency, depth and strength of ground-based inversions measured near Edmonton by minisonde appear in Table 1.5. Early-morning inversions are deepest in winter, averaging 467 m deep with a gradient of $0.035^{\circ}\text{C}/\text{m}$. Early-morning inversions are shallow in spring, averaging 275 m deep with a gradient of $0.017^{\circ}\text{C}/\text{m}$.

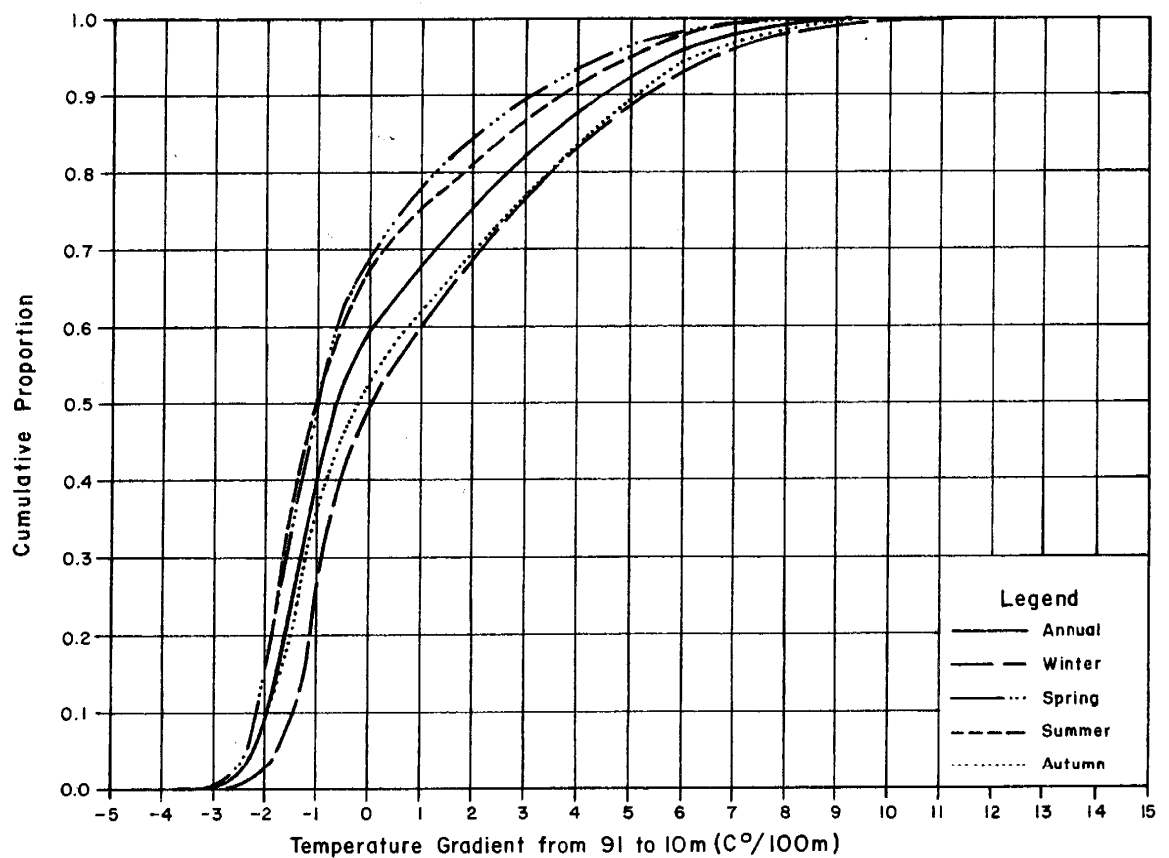
1.3 TURBULENCE

It is difficult to give a satisfactory definition of turbulence, although the "bumpiness" of an airplane ride and the "gustiness" of stormy weather are both familiar manifestations. This variability of the wind is generally conceived as arising from a mean motion with some fluctuations added. The mean motion is described by the laws of mechanics, but the fluctuations are not. The division of complicated motion into these two components is quite arbitrary, and really amounts to choosing the limit of useful detail. Turbulence is the part of the motions "containing details we shall not enquire into" (Scorer 1978). In practical terms it is only the overall effects of such motions that are important; hence statistical descriptions of the population are sufficient.

The problem of definition is but one of a vast number that are encountered in dealing with turbulent flow phenomena. It has been said that "the one uncontroversial fact

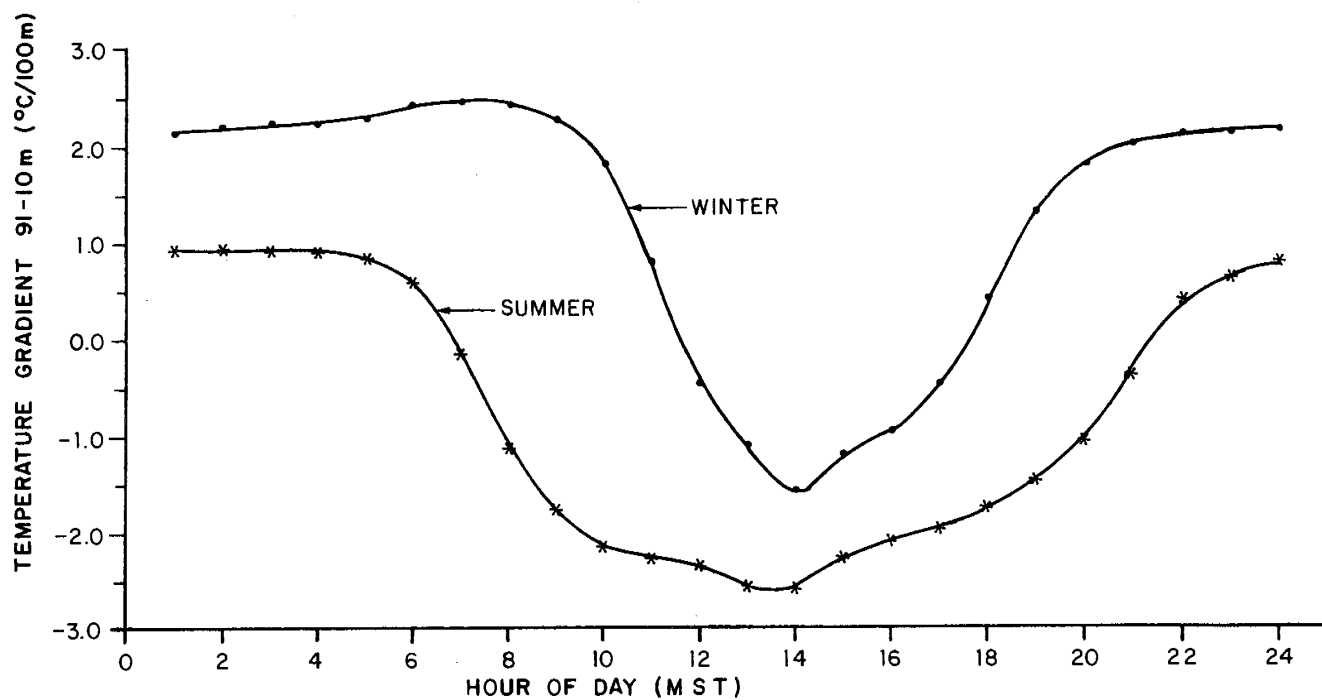


(a) Calgary Bonnybrook Tower

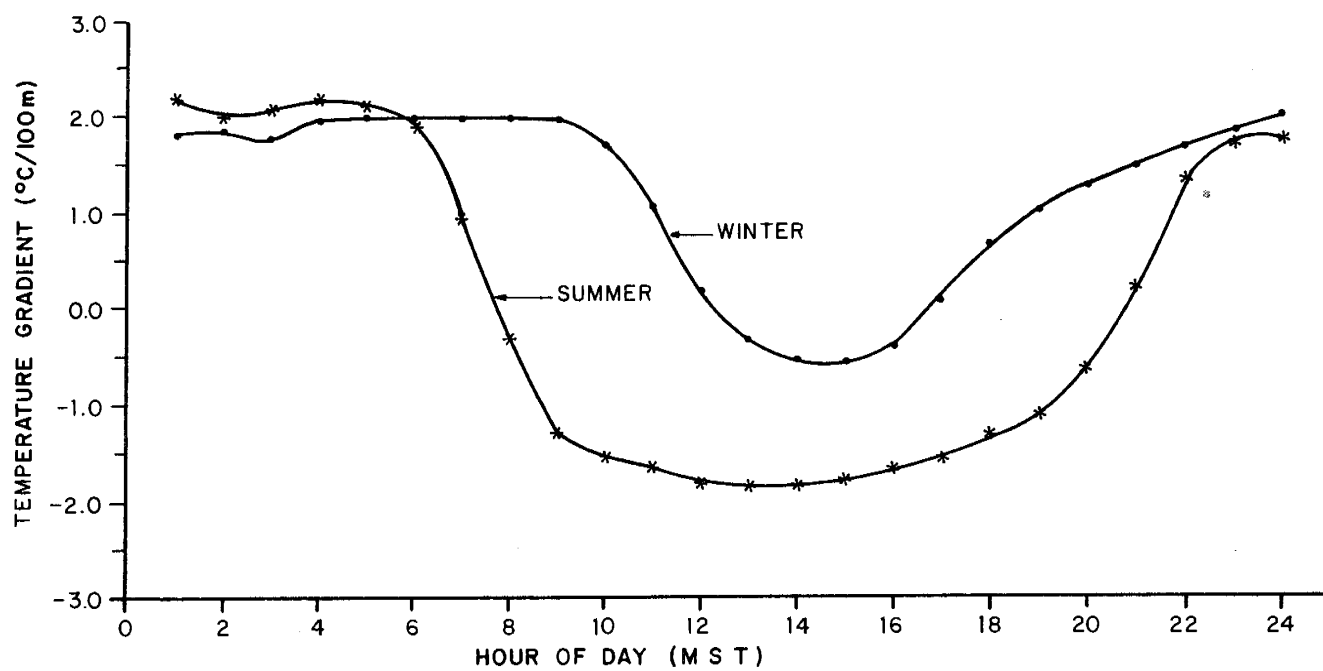


(b) Edmonton CBC Tower

Figure 1.23 Cumulative frequency distributions of temperature gradients from 91 to 10m, 1974 - 1980.



(a) Calgary Bonnybrook Tower



(b) Edmonton CBC Tower

Figure 1.24 Diurnal variation of temperature gradient.

Table 1.2 Joint frequency distribution of temperature gradient with wind direction at 91 m for the Edmonton CBC Tower, 1974 - 1980 (Choukalos 1985).

Wind direction to 8 points	TEMPERATURE GRADIENT (°C/100 m)							TOTAL
	< or = -1.90	-1.90 <TG<= -1.70	-1.70 <TG<= -1.50	-1.50 <TG<= -0.50	-0.50 <TG<= 1.50	1.50 <TG<= 4.00	>4.00	
Calm	0.25	0.10	0.09	0.46	0.26	0.28	0.27	1.71
N	1.75	1.02	1.24	5.89	2.05	0.89	0.49	13.13
NE	0.53	0.38	0.39	1.91	1.02	0.51	0.21	4.93
E	0.89	0.39	0.52	2.85	1.39	0.84	0.37	6.84
SE	1.40	0.87	0.88	4.47	3.22	1.66	0.95	13.44
S	1.29	0.97	0.83	2.91	3.41	3.01	2.93	15.35
SW	0.82	0.48	0.34	1.21	1.45	2.20	2.89	9.18
W	1.38	0.89	0.84	2.82	3.30	4.48	3.19	16.90
NW	2.09	1.28	1.39	6.86	3.66	2.01	1.27	18.54
TOTAL	10.21	6.31	6.51	28.97	19.75	15.88	12.38	100.00

Table 1.3 Joint frequency distribution of temperature gradient with wind direction at 10 m for the Calgary Bonnybrook Tower, 1974 - 1980 (Choukalos 1985).

Wind direction to 8 points	TEMPERATURE GRADIENT (°C/100 m)							TOTAL
	< or = -1.90	-1.90 <TG<= -1.70	-1.70 <TG<= -1.50	-1.50 <TG<= -0.50	-0.50 <TG<= 1.50	1.50 <TG<= 4.00	>4.00	
Calm	0.98	0.14	0.19	1.69	4.32	5.72	5.38	18.40
N	2.30	0.63	0.77	4.94	2.92	1.38	0.65	13.58
NE	1.39	0.33	0.40	3.38	1.55	0.63	0.35	8.03
E	2.31	0.51	0.63	4.21	1.61	0.73	0.35	10.68
SE	2.29	0.28	0.39	2.39	1.42	0.64	0.34	7.75
S	2.63	0.36	0.44	3.08	1.59	0.74	0.58	9.44
SW	1.40	0.24	0.24	1.77	1.68	0.75	0.61	6.36
W	2.85	0.53	0.52	3.14	3.69	1.88	0.97	13.58
NW	2.39	0.44	0.51	3.66	3.25	1.59	0.65	12.50
TOTAL	18.55	3.47	4.10	28.27	21.71	14.08	9.85	100.00

Table 1.4 Mean 18-m to 8-m temperature gradient ($^{\circ}\text{C}/100\text{m}$) at Ellerslie,
1979 - 1985 (Sakiyama and Wong 1987).

Hour	March	June	September	December
1	3.22	2.33	2.44	7.48
2	2.18	2.03	2.35	7.10
3	2.98	2.32	2.10	7.05
4	3.34	2.50	1.54	6.48
5	3.18	2.39	1.41	6.88
6	2.73	1.58	0.73	6.85
7	2.74	-1.71	0.24	6.81
8	1.38	-4.34	0.08	6.47
9	-0.12	-5.11	-3.74	5.45
10	-3.26	-5.25	-5.83	4.82
11	-5.93	-5.47	-6.67	3.49
12	-6.99	-4.94	-7.09	1.43
13	-7.42	-4.85	-7.33	-0.42
14	-7.56	-4.95	-7.44	-0.97
15	-7.43	-4.75	-7.39	-0.81
16	-7.18	-4.52	-6.85	-0.14
17	-6.40	-3.71	-6.31	1.25
18	-5.38	-4.03	-5.05	3.16
19	-3.61	-2.99	-3.31	5.76
20	-1.04	-1.72	0.32	6.61
21	1.08	-0.09	2.28	7.35
22	2.18	1.95	4.21	7.66
23	2.53	3.01	3.71	7.82
24	2.68	3.62	2.81	8.06

Table 1.5 **Summary of inversion characteristics measured by minisonde near Edmonton, 1977 - 1982.**

Season	Time	Frequency of occurrence (%)	Mean Depth in m (max)	Mean Temperature Gradient in °C/100 m (max)
Fall	07:30	47	382 (895)	2.6 (5.6)
	09:00	37	369 (1125)	2.2 (5.0)
	12:00	5.6	250 (612)	1.5 (2.3)
	15:00	7.9	88 (414)	1.2 (2.3)
Winter	09:00	33	467 (1350)	3.5 (13.4)
	12:00	19	396 (990)	2.7 (5.5)
	15:00	21	350 (1080)	2.2 (4.6)
Spring	07:30	58	275 (706)	1.7 (4.3)
	09:00	23	151 (612)	1.3 (3.2)
	12:00	9.1	90 (216)	1.8 (3.5)
	15:00	7.8	66 (108)	1.2 (2.1)
Summer	06:00	75	403 (1080)	2.2 (4.9)
	09:00	3.0	24 (36)	1.2 (1.5)
	12:00	4.0	45 (108)	1.0 (1.1)

about turbulence is that it is the most complicated kind of fluid motion." Although agreement is not universal, turbulence will generally be described as (e.g., Panofsky and Dutton 1984):

- (1) Irregular - The motions are irregular with apparently random variations in space and time. The behavior of fluid parcels is chaotic and disorganized.
- (2) Diffusive - If one follows a marked fluid parcel it will be observed to wander about, making excursions farther and farther from its initial location. This transport of the fluid and any characteristic attached to it (such as heat or chemical composition) results in irreversible mixing. Although qualitatively similar to molecular action, the rate of transfer is several orders of magnitude larger than that due to molecular diffusion. A flow may seem irregular, but if it does not exhibit spreading of fluid characteristics then it is not turbulent. Waves, for example, can cause fluid parcels to oscillate in a seemingly random fashion, but when the wave ceases the parcels return to their mean positions. No mixing has occurred; therefore the motion is not turbulence. The diffusivity of turbulence is its most important property as far as applications are concerned, so much so that turbulence has been defined as "a complex motion which causes diffusion...."
- (3) Operating at Large Reynolds Number - The Reynolds number is the ratio of inertia forces (mass $\times$ acceleration) to viscous forces. For turbulence to occur, the viscous forces must be negligible compared to the inertia forces. Cream is easily stirred into coffee as the result of turbulence produced by the spoon; however, the same spoon movements will not readily homogenize house paint.
- (4) Dissipative - Energy is transferred from larger motions to smaller and smaller motions until finally the kinetic energy of the fluid motion is transformed into internal energy of the fluid. Heat is produced as work is done against viscosity. Random sound waves (noise) may be irregular and dispersive, but they are not dissipative and hence not turbulence. To make up for viscous losses, turbulence must have a continuous supply of energy; otherwise it will decay.

- (5) **Rotational** - Fluid parcels or eddies possess angular velocity, one half of which is called vorticity. A tube of vorticity can be associated with these eddies. Turbulence has also been defined as "chaotic vorticity" (Scorer 1978). Random waves on a water surface are not turbulence because they are irrotational.
- (6) **Three-Dimensional** - The interaction of eddies leads to the stretching of vortex tubes, reducing the cross-sectional area and increasing the local vorticity in accordance with conservation of angular momentum. This amplification of vorticity is another aspect of the energy transfer from large motions to smaller motions. The vortex stretching mechanism can occur only in a three-dimensional flow field. Weather systems (cyclones and anticyclones) are essentially two-dimensional and are not considered to be "ordinary turbulence". They can, however, be described by similar statistical techniques and are sometimes viewed as two-dimensional turbulence.
- (7) **Part of a Continuum** - Continuum concepts are applicable because the smallest motions are much larger than molecular distances or intermolecular spaces.
- (8) **A Feature of the Flow** - Turbulence does not refer to a property of a fluid, but rather of a fluid motion. Turbulent flows are governed by initial conditions as well as boundary conditions. Local properties are not necessarily related to local conditions. The time and length scales of the turbulent motion are as large as the scales of the distributions of the properties being transported. This is in sharp contrast to molecular motions where the time and length scales are very much smaller than the scales of the property being transported.

1.3.1 Origins of Turbulence

Turbulence arises as the result of instabilities in an initially laminar flow. In laminar flows adjacent layers of fluid remain distinct and identifiable and the paths of fluid parcels are parallel. In turbulent flows, no layering can be discerned and parcel paths are intermingled in a chaotic fashion. The details of the transition from laminar to

turbulent flow are not well understood. The criteria for transition are not exact, but usually related to some threshold value of the Reynolds number, defined by:

$$\text{Re} = (\text{a characteristic flow length}) \\ \times (\text{a characteristic flow velocity}) \\ \div (\text{dynamic viscosity/density})$$

Laboratory flows with Reynolds numbers greater than about 1000 are usually turbulent.

Atmospheric flows exhibit great variability in characteristic velocities and lengths, a variability complicated by the effects of temperature stratification. Consequently the transition from laminar to turbulent flow and the reverse are common in the atmosphere above the boundary layer (Scorer 1978), though they are gradual, taking place over many minutes or hours.

Once initiated, turbulence cannot maintain itself against the viscous dissipation unless there is some source of energy. Two common energy sources are:

- (1) shear of the mean wind (spatial variations of the average wind speed)
- (2) buoyancy (partially organized upward motion of heated air parcels).

Wind shear is usually the result of frictional retardation near solid boundaries. Ground cover causes wind speed changes in the vertical, while large obstacles such as buildings can also cause wind speed changes in the horizontal. Both situations allow turbulence to extract energy from the mean airflow. The largest eddies are about the size of the region in which the shear exists, for example, a building produces eddies not much larger than its largest dimension. The turbulence arising from shear is known as *mechanical turbulence* and the surface protrusions (grass, crops, stones, forests) causing it are termed *roughness elements*. Mechanical turbulence increases with wind speed and with terrain roughness. Terrain roughness is characterized by a "roughness length" which represents the size of eddies that can exist among the roughness elements. Thus the size, shape and spacing of the roughness elements are important in determining the roughness length. The wind speed fluctuations associated with mechanical turbulence have periods that are measured in seconds.

Buoyancy usually results from air parcels becoming warmer than their surroundings through contact with ground warmed by the sun. These warm parcels rise and eventually organize themselves into thermals, which may rise to several hundred metres upward from the surface. This process is termed convection and is accompanied by a superadiabatic temperature gradient near the ground. The rising of the thermals and the motion of the cooler air descending to take their place produces *thermal turbulence*. Because of the dependence on solar heating thermal turbulence is a function of such variables as latitude, season, time of day, altitude, cloudiness, surface albedo and surface heat properties. Other heat sources such as cities and volcanoes also produce thermal turbulence. The fluctuations associated with thermal turbulence have periods that are measured in minutes.

1.3.2 The Cascade of Energy in Turbulence

Turbulence in the atmospheric boundary layer consists of eddies ranging in size from a few centimetres to several hundreds of metres. This wide size range is attributed to a breakdown mechanism whereby large eddies become unstable and transfer their energy to smaller eddies. These smaller eddies, in turn, become unstable and generate still smaller eddies. This process continues until the kinetic energy is dissipated into heat by viscosity. This "energy cascade" is shown schematically in Figure 1.25. This physical picture of developed turbulence is succinctly described in the often-quoted verse (L.F. Richardson):

Big whorls have little whorls,
Which feed on their velocity;
And little whorls have lesser whorls,
And so on to viscosity
(in the molecular sense).

In terms of vorticity, turbulence is viewed as a tangle of vortex tubes that are stretched in a preferred direction by the mean flow and in a random direction by one another. As the vortex tube is stretched its cross-sectional area decreases and its kinetic energy of rotation increases at the expense of the kinetic energy of the velocity component responsible for the stretching. The increased velocity in the perpendicular plane then stretches vortices aligned in that direction and so on. Each extension of the tube in one direction decreases the size and increases the velocity of motions in the other two

directions. The net result is a progressive transfer of kinetic energy to smaller and smaller scales until it is finally converted into internal (heat) energy by the viscous action on the smallest motions.

A graph of the energy associated with an eddy as a function of eddy size is called a spectrum (Figure 1.26). Energy enters the spectrum in large motions and is cascaded through smaller and smaller motions until it is dissipated. With each breakdown into smaller eddies, the orienting effect of the mean flow is weakened. For sufficiently small eddies there is no preferred direction; the small eddies have no "memory" of the way in which the energy was fed into the turbulence. This portion of the spectrum is governed by only two parameters -- the mean rate of energy dissipation and the kinematic viscosity. This is known as the *equilibrium range*. The largest eddies that fit into this range are small compared to the height above ground. Viscosity begins to annihilate eddies smaller than the Kolmogorov microscale which is in the order of 1 mm for the atmospheric boundary layer. At the large size end of the equilibrium or local isotropy range, viscosity has little effect and the spectrum depends only on the rate of transfer of energy. This region, where there is no production or dissipation of energy, is called the *inertial subrange*.

1.3.3 Statistical Description of Turbulence

The chaotic motions that comprise turbulence make deterministic predictions difficult. The state of the fluid at a given time will be unique and will represent but one of many possible states that could have occurred. It is therefore appropriate to adopt statistical methods that allow the prediction of the probability that a certain state will occur. The theory of turbulence is heavily dependent on the theory of statistical random processes.

A process is said to be *stationary* or *steady* if the statistical properties for a time series do not vary with time. Thus the mean and standard deviation of the wind speed for one 10-minute interval would be similar to that for any other 10-minute interval. Both diurnal and synoptic weather cycles cause major changes in the atmospheric boundary layer. The condition may be fulfilled approximately for intervals short compared with the diurnal cycle, that is, for a few hours (when the large-scale pattern does not change).

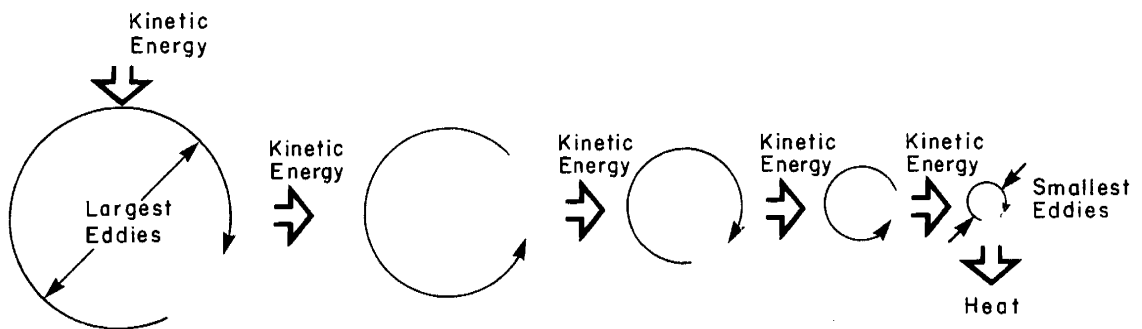


Figure 1.25 The energy cascade.

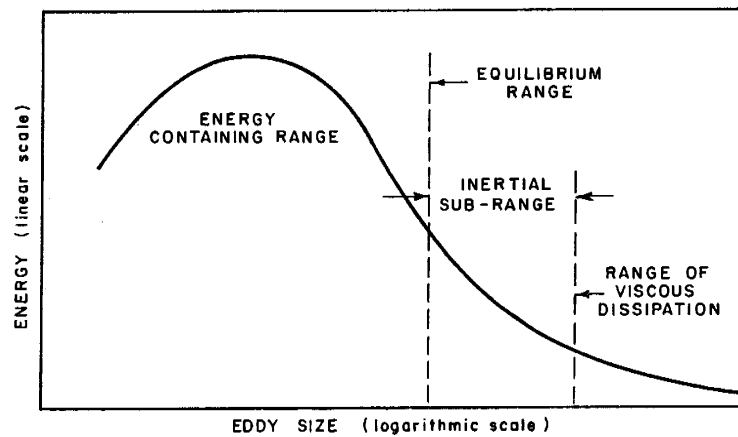


Figure 1.26 Spectrum of turbulence.

A process is said to be *homogeneous* when its statistical properties do not vary in space. A measurement at a particular point will display statistical characteristics identical to those taken at some neighboring point. Horizontal homogeneity is a reasonable assumption over regions of similar surface characteristics. Vertical homogeneity cannot, in general, be achieved because of the presence of the earth's surface and the action of the forces of buoyancy and gravity.

A process is said to be isotropic when its statistical properties are independent of direction. Owing to wind shear, the atmospheric boundary layer cannot, in general, be isotropic. However, as described earlier, eddies in the equilibrium size range will be isotropic, a condition which is usually termed *local isotropy*.

A single observation of the outcome of a process is called a *realization*. A grouping of large number of realizations is known as an ensemble. Statistical laws are derived for *ensembles*. For single realizations all that can be predicted is the probability of a certain outcome.

A process is said to be *Gaussian* if the frequency distributions of its measurable characteristics follow the familiar normal, bell or gaussian curve. Such distributions are completely specified by the mean and standard deviation. Many variables in the atmospheric boundary layer have been found to be approximately Gaussian in distribution.

Turbulent flows are generally decomposed into a *mean* component and a *fluctuating* component. The division is a choice which may be completely arbitrary, a matter of convenience or based on some natural aspect of the flow such as geometry. By convention the Cartesian coordinate system is oriented so that the x-axis lies along the mean wind direction, the y-axis lies perpendicular to the mean wind direction and the z-axis extends vertically upward from the ground. The instantaneous wind velocity then has three components u , v and w in the x , y and z directions respectively, which can be written as:

$$u = u_{avg} + u'$$

$$v = v_{avg} + v' \quad (v_{avg} = 0 \text{ by virtue of the axis orientation})$$

$$w = w_{avg} + w' \quad (w_{avg} = 0 \text{ for horizontal flows})$$

where "avg" represents the mean components and the primes represent the fluctuations. These deviations from the mean may be positive or negative. They form a time series that can be subjected to various types of statistical analysis.

Turbulence is most often described quantitatively by the standard deviations.

$$\sigma_u = \sqrt{(u')^2 \text{ avg}}$$

$$\sigma_v = \sqrt{(v')^2 \text{ avg}}$$

$$\sigma_w = \sqrt{(w')^2 \text{ avg}}$$

The *intensity* of turbulence is given by the ratio of component standard deviation to mean speed:

$$i_x = \sigma_u/u_{\text{avg}} \quad i_y = \sigma_v/u_{\text{avg}} \quad i_z = \sigma_w/u_{\text{avg}}$$

These intensities take values in the order of 0.1 during the day. Steenbergen (1971) reported turbulence intensities measured in a rural area south of Edmonton in September as $i_y = 0.16$, $i_z = 0.08$, $i_x = 0.19$ with corresponding intensities in the city of Edmonton a factor of 3 to 4 higher.

Wind fluctuations can be regarded as arising from the superposition of numerous sinusoidal waves. Each such wave is characterized by:

- (1) a period T (s), the time required for one complete oscillation or cycle,
- (2) a frequency n (Hz), the number of complete cycles per unit time, $n = 1/T$,
- (3) a wavelength λ (m), the distance between wave crests or points in the same phase of oscillation,
- (4) an angular frequency ω ($\text{rad} \cdot \text{s}^{-1}$), the number of radians of rotation in unit time; $\omega = 2\pi/T$,

- (5) a wavenumber k (m^{-1}), the number of waves in unit length; $k = 1/T$.
- (6) an angular wavenumber κ ($\text{rad} \cdot \text{m}^{-1}$), the number of radians of rotation in unit length.

It is very difficult to measure eddy size directly; most measurements are made at a fixed point in space over a period of time. Because eddies interact, grow and decay with time such a time series may not be simply related to the distribution of eddy sizes. If the turbulence is changing slowly enough, time variations observed at a fixed point will be nearly identical to the spatial variations observed at a given instant of time. In other words the turbulence is "frozen" into the flow and the wind merely advects the eddies past the observation point. If such is the case, the frequency and wavelength of each eddy are related by the usual wave equation $n = u/\lambda$ where u is the mean wind, now also the phase or translation speed of the waves. Taylor's "frozen turbulence" hypothesis is found to be a good approximation for small eddies and weak turbulence in the atmospheric boundary layer. In strong turbulence, large eddies (wavelength greater than 100 m) apparently do not move at the average speed of the wind.

The fluctuations of the wind can also be described in terms of the correlation between values at different times. The *autocorrelation coefficient* is defined by:

$$R(\tau) = [u'(t) u'(t+\tau)]_{\text{avg}} / (u'^2)_{\text{avg}}$$

where $u'(t)$ is the fluctuation at time t , $u'(t + \tau)$ is the fluctuation at a time τ later, and u'^2 is the variance over a suitable time interval.

The product of velocity fluctuations is calculated for different values of τ (the "lag" time) and the average is then taken. The variance is assumed constant over the interval of the averaging (stationary turbulence). The value of R is close to 1 for short lag times and is expected to decrease towards zero at large lag times (Figure 1.27). The shape of the curve $R(\tau)$ versus τ will depend on the structure of the turbulence. If only small eddies are present, the autocorrelation coefficient decreases more rapidly than if large

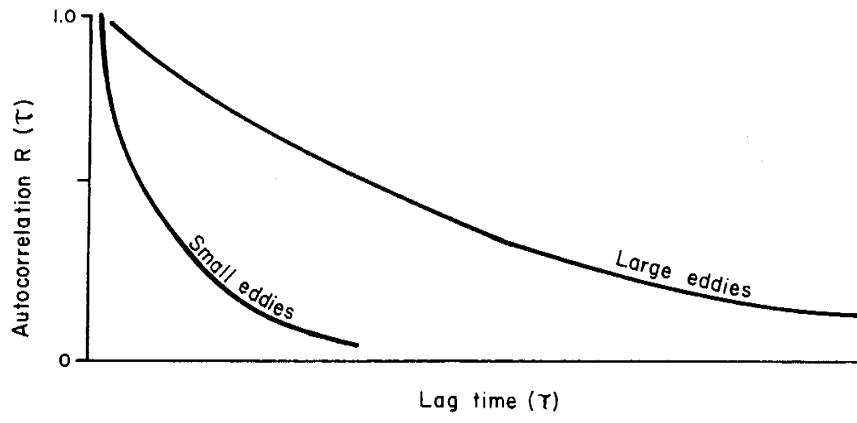


Figure 1.27 Autocorrelation curve.

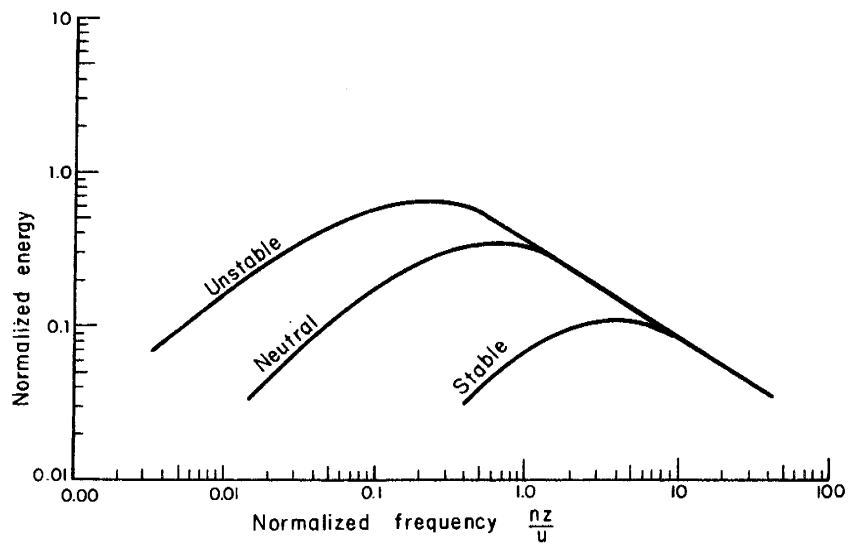


Figure 1.28 Conventional presentation of vertical wind spectra.

eddies are present. The area under the curve is therefore a measure of the period associated with the largest eddies in the turbulence. The *integral time scale* of turbulence is thus defined by:

$$T_E = \int_0^{\infty} R(\tau) d\tau$$

Assuming the frozen turbulence the size of the largest eddies is $L_E = uT_E$, which is also known as the *integral length scale* of turbulence.

Since the variance of velocity fluctuations is twice the turbulent kinetic energy per unit mass of air, splitting up the total variance into contributions from different sizes of eddies produces an energy spectrum (Kaimal et al. 1968). This is done by Fourier analysis, a technique based on the principle that any time-dependent variable can be represented by the sum of sine and cosine waves of different frequencies and amplitudes (Pasquill 1974). Each such wave is known as a Fourier component. As might be expected, the energy spectrum and the autocorrelation are related, in fact the autocorrelation coefficient and the power spectral density are Fourier transforms of one another.

Spectra are generally plotted with normalized energy or variance on a logarithmic ordinate scale as a function of reduced frequency nz/u on a logarithmic abscissa scale (Figure 1.29). The use of a dimensionless, (normalized) frequency makes the spectra nearly independent of wind speed and height near the ground. Two parameters can be used to characterize the general shape and position of the spectra:

- (1) The wavelength λ_m (or frequency n_m) corresponding to the peak in the spectrum: This defines the size of the "most energetic" eddies and serves as a practical scale of turbulence. Its relationship to the integral scale depends on the shape of the spectrum. For the vertical wind component in neutral conditions, λ_m is approximately four times the integral length scale and about twice the height. The horizontal components generally contain more energy at larger wavelengths. For the v-component, λ_m is typically 150 to 300 m in neutral conditions showing no obvious dependence on height. For the u-component, λ_m appears to increase slowly with height taking values of several hundred metres.

- (2) The rate of energy dissipation ϵ , which corresponds to the rate of energy transfer between eddy sizes in the cascade process and defines the spectrum in the inertial subrange. Values of ϵ may be as large as $10 \text{ m}^2\text{s}^{-3}$ close to the ground and vary inversely with height in neutral conditions. In unstable conditions the height dependence virtually disappears.

If a spectrum extends over a broad frequency range, the existence of any local minimums will be an important attribute. In the atmosphere it is sometimes held that such a spectral "gap" is present at a frequency of about one cycle per hour (0.003 Hz). It is then possible to represent a time series as the sum of two random processes, one slowly varying (synoptic weather systems) and the other rapidly varying (ordinary turbulence). Sometimes the former is called the *macrometeorological* regime and the latter the *micrometeorological* regime. It also becomes possible to determine an optimum ("natural") averaging period such that the mean and variance will not change significantly as the period is lengthened. It now appears that the existence of the "gap" is a rather idealized situation. Wherever there are local airflows such as those induced by topography, there is likely to be energy in what otherwise would be the "gap". This is sometimes called the *mesometeorological* regime.

Various spectra observed by Walker (1964) over the rolling prairie at Suffield, Alberta are shown in Figure 1.29. There is a macroscale peak at 0.015 cycles per hour (4-day period). The flat low-energy region between 1 and 10 cycles per hour represents a spectral gap.

Most atmospheric measurements are obtained from points fixed in space, in what is called the *Eulerian* reference system. The alternative is to follow the behavior of a single air parcel as it moves through the atmosphere. When the coordinates are attached to a moving parcel, a *Lagrangian* reference frame is being used. Although it is difficult to make Lagrangian measurements, much of fluid mechanical theory actually refers to such a moving system. The relationship between the two systems is of obvious importance. A Lagrangian integral time scale T_L and Lagrangian length scale can similarly be defined. It is often assumed that the autocorrelations are related by $\beta T_L = T_E$ where β is a constant ranging from 1 to 10. This relation reflects the intuition that variations are slower in the Lagrangian reference frame.

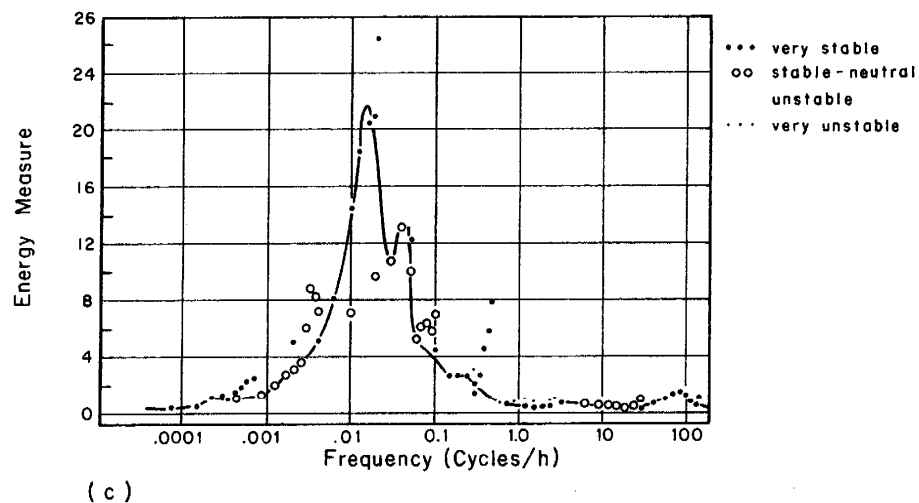
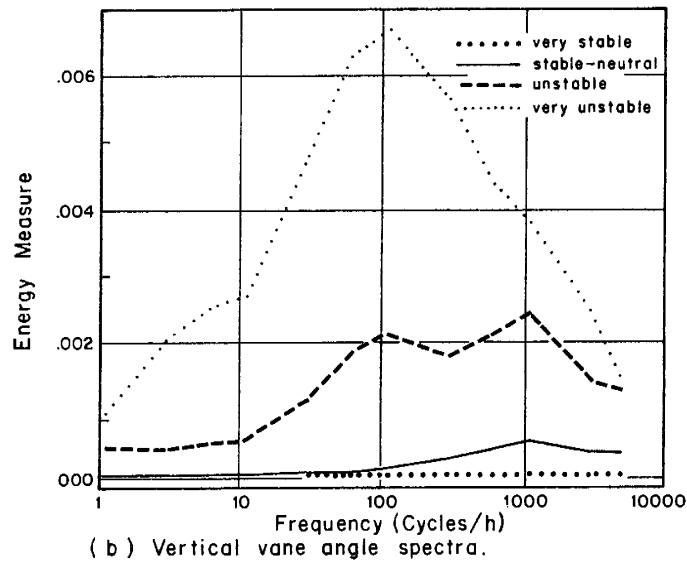
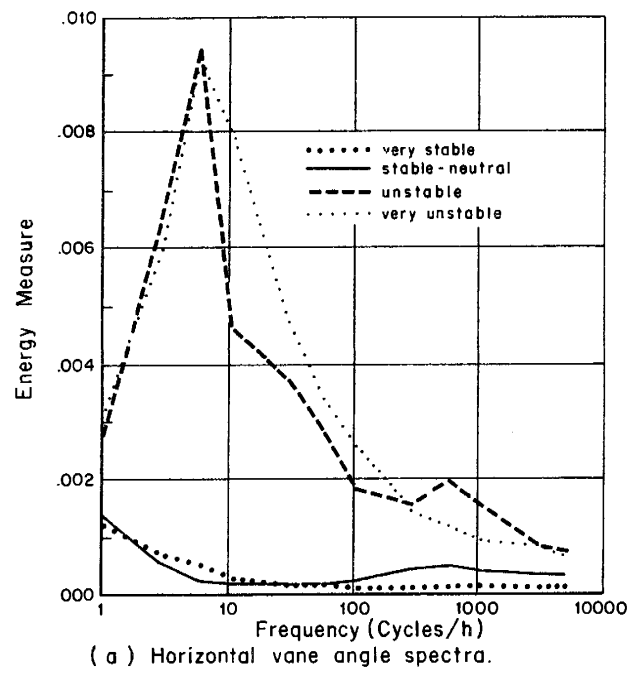


Figure 1.29 Wind spectra at Suffield (Walker 1964).

1.3.4 The Effect of Buoyancy on Turbulence

In the atmospheric boundary layer, energy is supplied to turbulence through the working of the horizontal shear stresses. Temperature stratification gives rise to buoyancy forces which either augment or diminish this energy, depending upon the direction of heat transfer.

The shear stress τ , which is the force per unit horizontal area required to maintain the flow, is equal to the downward flux of momentum to the ground, that is,

$$\tau = -\rho (u'w')_{\text{avg}}$$

where ρ is the density of air and $(u'w')_{\text{avg}}$ is the covariance of the vertical and longitudinal wind fluctuations over a suitable time interval.

The stress per unit density turns out to be convenient parameter and, since it has the dimensions of velocity squared, the so-called *friction* or *shear stress velocity* is defined as

$$u_* = (\tau/\rho)^{1/2}$$

Shear stress is approximately constant in the first few tens of metres above the ground, a region known as the *constant stress* or *surface stress layer*. It then decreases steadily to zero at the top of the friction layer (planetary boundary layer or Ekman layer). By convention the symbols always refer to surface values unless otherwise stated.

Ignoring the heat contained in water vapor, the sensible heat flux is given by

$$H = \rho c_p (w'T')_{\text{avg}}$$

where ρ is the density of air, c_p is the specific heat of air at constant pressure and $(w'T')_{\text{avg}}$ is the covariance of the vertical wind and temperature fluctuations over a suitable time interval.

During the day, the sun warms the ground and heat is transferred upward, away from the earth's surface. This heat flux is taken to be positive, consistent with the positive direction for the axis and the vertical wind component w . This buoyancy adds energy to the turbulence. At night, the ground cools by radiation to space and the heat flux is reversed. In this situation energy is extracted from the turbulence. Heat flux varies from about -30 W/m^2 to $+300 \text{ W/m}^2$.

The energy per unit time per unit mass of the buoyancy forces is given by

$$B = gH/(\rho c_p T) = g/T (w' T')_{\text{avg}}$$

where T is the average air temperature and g is the acceleration due to gravity.

The energy per unit time per unit mass of the shear stress is given by

$$S = \tau/\rho \, du/dz = (u' w')_{\text{avg}} \, du/dz$$

where du/dz is the gradient of the longitudinal wind.

The *flux Richardson number* Rf is then defined as the ratio between the consumption of turbulent energy by buoyancy forces and the production of turbulent energy by shear stress, that is, the negative of the ratio of the two production terms:

$$Rf = -B/S$$

When Rf is negative, both buoyancy and shear stress are adding energy; existing turbulence tends to increase and the atmosphere is said to be *unstable*. When Rf is near zero, only shear production is present; existing turbulence is unaffected and the atmosphere is said to be *neutral*. When Rf is positive, buoyancy is removing energy; existing turbulence tends to decrease and the atmosphere is said to be *stable*. It is expected that at some positive value of Rf the consumption of energy by buoyancy forces becomes so large that existing turbulence disappears altogether. This critical flux Richardson number is likely less than 1.0 because a large portion of shear-produced energy is "cascaded" down the spectrum to be dissipated into heat. The value seems to lie around 0.2 but may depend upon the circumstances.

The flux Richardson number is a function of height above ground, even in the constant stress layer, due to the height dependence of the wind shear. Above the surface layer the shear stress and heat flux both decrease with height as well; consequently R_f varies in some non-linear fashion. At large heights above the ground or with large heat flux, shear production becomes less and less important, until at some level the flow is essentially buoyancy driven. In other words, forced convection gives way to free convection. The flux Richardson number is a measure of the effect of local heat flux upon turbulence. It has no global significance like the Reynolds number.

It is often assumed that fluxes are proportional to vertical gradients:

$$(-u'w')_{\text{avg}} = K_m du/dz$$

$$(-w'T')_{\text{avg}} = K_h d\theta/dz$$

where K_h is the eddy diffusivity of heat, K_m is the eddy diffusivity of momentum and $d\theta/dz$ is the potential temperature gradient (to allow for the compressibility of air).

Substitution in the definition of R_f yields

$$R_f = (K_h / K_m) Ri$$

where the *gradient Richardson number* has been defined as

$$Ri = \frac{(g/T) d\theta/dz}{(du/dz)^2}$$

This dimensionless parameter has the advantage of being expressed in terms of vertical profiles which are much more easily measured than are the fluxes. However, the gradient Richardson number has a different meaning. It indicates the stability with respect to small external disturbances of a stratified atmosphere initially without turbulence. In other words, it is a measure of the onset of turbulence. If Ri in an initially laminar flow is smaller than some critical value Ri_{crit} , then turbulence is likely to occur, if it is larger than Ri_{crit} , then turbulence cannot get started. In the absence of viscosity this critical value is found to be 0.25. Once turbulence has come into existence it will remain until the flux Richardson number exceeds the critical value for suppression. It is not clear how $R_{f_{\text{crit}}}$ and Ri_{crit} are related, for turbulence has been

observed at practically all gradient Richardson numbers (Plate 1971). Recent evidence seems to support Richardson's original suggestion that turbulence is damped for $Ri > 1$. It is entirely possible that the value depends upon the structural details of the turbulent flow.

Whenever the gradient Richardson number lies between these upper and lower critical values the state of the atmosphere is determined by its past history: laminar flow remains laminar until Ri drops below the lower critical value and turbulent flow remains turbulent until Ri exceeds the upper critical value. This Richardson number hysteresis is the mechanism by which turbulent interfaces in stratified flow are kept well defined and by which inversion layers thicken.

The gradient Richardson number indicates that the nature of turbulence depends on both thermal stability and wind shear. Turbulence can be suppressed only when the potential temperature gradient is positive, but that alone does not ensure damping. It is quite possible for turbulence to be maintained in a temperature inversion by the action of wind shear. In the terminology of deductive logic, thermal stability is a necessary, but not sufficient condition for the suppression of turbulence.

Near the ground wind shear is generally large, so ground-based inversions are likely to contain mechanical turbulence. Because the shear decreases with height there may exist a level h_{crit} where the gradient Richardson number falls below Ri_{crit} . Above this level the atmosphere is turbulence-free, whereas below, the atmosphere is in a state of damped mechanical turbulence.

In summary, the gradient Richardson number can be used as an index of stability, with the meaning given below (Panofsky 1969):

$Ri < -0.04$	thermal turbulence dominates
$-0.03 < Ri < 0$	thermal and mechanical turbulence
$Ri = 0$	mechanical turbulence only
$0 < Ri < Ri_{crit}$	mechanical turbulence weakened by stratification
$Ri > Ri_{crit}$	turbulence suppressed

Although the gradient Richardson number has been used extensively, it has a number of serious limitations as a representation of the effects of buoyancy. Current knowledge

supports the use of the *Monin-Obukhov length*, defined by

$$L = \frac{-u_*^3}{k(g/T)(w'T')_{\text{avg}}}$$

where k is the von Karman constant (added for convenience) which has a value of about 0.4. The quantity L is the height above ground where buoyant production of turbulence equals mechanical production. Thus L is independent of height and can serve as a useful length scale. The stability parameter for the surface layer is then simply the ratio of height to Monin-Obukhov length (z/L) and the following may be stated (Holtslag 1984):

$z/L < -0.83$	very unstable
$-0.83 < z/L < -0.25$	moderately unstable
$-0.25 < z/L < -0.05$	slightly unstable
$-0.05 < z/L < 0.05$	neutral
$0.05 < z/L < 0.10$	slightly stable
$0.10 < z/L < 0.25$	moderately stable
$0.25 < z/L < 1$	very stable
$z/L > 1$	extremely stable

An immediate inference that can be drawn is that the nature of turbulence will change with height. Close to the ground (small z) the atmosphere is always near neutral (mechanical turbulence dominant). At height L the buoyancy forces are as important as the shear forces. Above L buoyancy effects dominate.

The three possible stability parameters for the surface boundary layer are related by

$$R_f = \frac{K_h}{K_m} \quad Ri = \frac{z}{L\phi_m}$$

where ϕ_m is a function describing the wind shear. Of these z/L is widely preferred as the most basic. L is independent of height whereas both R_f and Ri may vary in complicated ways. In near-neutral conditions all three tend to be equal. In unstable conditions $Ri \approx z/L$ and in stable conditions Ri is a variable fraction of z/L .

In the outer portion of the atmospheric boundary layer, the Coriolis force becomes important. A suitable parameter is the Kazanski - Monin stability. (Smith 1979, Tagliazucca and Nanni 1983).

$$\mu = \frac{ku_*}{fL}$$

where k is von Karman's constant and f is the Coriolis parameter. The depth of the boundary layer in neutral conditions is proportional to u_*/f , so μ represents the ratio of neutral boundary layer height to the Monin-Obukhov length. The stability of the entire boundary layer is then (Schayes 1982):

$\mu > 20$	very stable
$10 < \mu < 20$	stable
$-20 < \mu < -20$	neutral
$-40 < \mu < -20$	unstable
$\mu < -40$	very unstable

Under strong solar heating, large thermal eddies become organized into well defined updrafts and downdrafts. Turbulence is found to scale with the ratio of the convective mixed-layer height, h , to Monin-Obukhov length and is categorized as (Briggs 1988):

$-h/L > 100$	very unstable
$20 < -h/L < 100$	moderately unstable
$5 < -h/L < 20$	slightly unstable
$-h/L < 5$	near neutral

1.3.5 Turbulence Typing Schemes

Fluxes and gradients are difficult and expensive to measure; in the early years, the technology to do so was not even available. Consequently the nature of turbulence was inferred either from simple on-site measurements or from large-scale meteorological observations. Much attention was focused on the use of the hourly observations made at aviation weather stations, the advantages being ready access to the data, broad spatial coverage and the existence of long periods of record. More recently, effort has been expended to devise simple schemes that do not require a human observer for use in site-specific environmental assessments. All typing schemes share the common goal of describing turbulence in terms of a small number of discrete classes that reflect the various combinations of mechanical and thermal components.

1.3.5.1 Energy Estimation Methods. Since turbulence near the ground is controlled by heat flux and wind shear, any related variables can be used to construct a typing scheme.

The oldest and most widely used typing scheme of this type was developed three decades ago (Pasquill 1961). This system was introduced as a practical substitute for measurements of turbulence, and was semiempirical in nature, based on the rather limited data available at the time. The construction of the stability categories implicitly represented two characteristic properties of airflow near the surface, namely, (a) the restriction of numerically small values of the Monin-Obukhov length L (or Richardson numbers greatly different from zero) to very light winds, and (b) the well-defined diurnal cycle of the vertical temperature profile in the first few metres above the surface and its dependence on the state of the sky (Pasquill 1974, p. 377 - 378). This scheme is shown in Table 1.6.

Such a simplified scheme is bound to be a rather crude specification of turbulence. Although the spread of fluctuation values within each class covers the entire range for all classes, the median values do change systematically from class to class (Luna and Church 1972, Ludwig and Dabberdt 1976). The limitations of the Pasquill scheme stem from the assumed links between solar radiation and heat flux and between wind speed and shear stress. Since the functional dependences of intervening variables are not included, the scheme represents but one of numerous possible relationships. Cloud opacity, surface albedo (fraction of incoming radiation reflected), soil heat transfer and evapotranspiration determine how much sensible heat flux results from a given solar input. The surface roughness determines the shear stress appropriate for a given wind speed. Therefore, the scheme can be expected to provide a good representation of turbulence only in situations where the assumed relationships hold. Where large departures are experienced, the scheme is unlikely to perform well.

Several modifications to the original Pasquill scheme have been proposed for operational programs. Insolation classes can be defined quantitatively (Hino 1968, Ludwig and Dabberdt 1976). Using solar altitude a completely objective method was devised (Turner 1961, 1964) so that an electronic computer could be used to calculate the frequency distributions using routine hourly observations at airport weather stations. The Atmospheric Environment Service uses this method to produce the so-called STAR (Stability Array) analysis for aviation weather stations. The frequency distributions for several Alberta airports are shown in Table 1.7. The results for any one climatic regime are likely to be quite similar. In the United Kingdom categories A and B account for about 10%, C and D for around 60%, E for about 10% and F for about 20% (Gifford

Table 1.6 Key to stability categories (Pasquill 1961).

Surface Wind Speed (m/s)	Insolation			Night Thinly overcast or $\geq 4/8$ low cloud	$\leq 3/8$ cloud
	Strong	Moderate	Slight		
<2	A	A-B	B	-	-
2-3	A-B	B	C	E	F
3-5	B	B-C	C	D	E
5-6	C	C-D	D	D	D
>6	C	D	D	D	D

(for A-B take average of values for A and B etc.)

Strong insolation corresponds to sunny midday in mid-summer in England, slight insolation to similar conditions in mid-winter. Night refers to the period from one hour before sunset to one hour after dawn. The neutral category D should also be used, regardless of wind speed, for overcast conditions during day or night, and for any sky conditions during the hour proceeding or following night as defined above.

A extremely unstable
 B moderately unstable
 C slightly unstable
 D neutral
 E slightly stable
 F moderately stable

Table 1.7 Frequency of occurrence of Pasquill stability classes at selected Alberta airports (STAR data).

Station	Pasquill Stability Category	Winter	Spring	Summer	Autumn	Annual
Calgary Int'l Airport 1972 - 1976	A	0.0	0.7	2.3	0.4	0.8
	B	1.9	6.9	12.5	6.0	6.9
	C	6.5	10.7	13.7	9.4	10.1
	D	58.0	55.8	46.1	47.7	51.9
	E	12.7	11.1	10.6	12.3	11.7
	F	20.9	14.9	14.8	24.2	18.7
Grande Prairie 1972-1976	A	0.0	0.9	1.3	0.2	0.6
	B	1.8	10.0	11.1	5.6	7.2
	C	8.7	12.5	13.0	9.1	10.8
	D	48.6	46.8	48.0	49.1	48.1
	E	9.9	10.6	9.2	9.2	9.7
	F	31.0	19.3	17.4	26.8	23.5
Medicine Hat 1972-1976	A	0.0	0.3	1.3	0.2	0.5
	B	1.5	5.4	10.6	4.3	5.5
	C	5.5	10.6	14.8	8.2	9.8
	D	59.8	55.8	44.3	51.6	52.9
	E	14.4	11.4	11.3	13.8	12.7
	F	18.8	16.5	17.7	21.9	18.7
Red Deer 1967-1976	A	0.0	0.3	1.1	0.1	0.4
	B	1.1	6.3	10.0	3.0	5.1
	C	6.0	11.2	15.1	8.9	10.3
	D	58.1	57.1	47.0	54.9	54.3
	E	18.9	13.4	12.7	18.2	15.8
	F	15.9	11.6	14.1	14.9	14.1
Edmonton Int'l Airport 1963-1972	A	0.0	0.4	1.2	0.0	0.4
	B	1.3	7.0	13.5	3.4	6.3
	C	6.7	11.8	15.9	9.0	10.9
	D	55.9	54.6	40.2	52.9	50.8
	E	14.2	12.5	10.4	14.3	12.8
	F	21.9	13.6	18.9	20.3	18.7

1976). This is not greatly different from the distributions in Alberta despite the climatic differences (maritime as opposed to continental). However, the U.K. is at approximately the same latitude as Alberta, so perhaps latitude is the overriding factor.

Another variation consists of deducing the heat fluxes from time of day, month and cloud amount or measured incoming solar radiation, and calculating the value of a continuous parameter P closely related to Pasquill's original categories (Smith 1972). This scheme is shown in Figures 1.30 and 1.31.

Schemes that use cloud amounts can only be used at aviation weather stations where a human observer is present. Schemes that use measured incoming solar radiation can only be used during the daylight hours and only during the snow-free portion of the year. Net radiation can be used to overcome these difficulties (Hayashi 1977, Williamson and Krenmayer 1980) with a classification such as that in Table 1.8.

A more simplified scheme (Briggs 1973) has been devised for applications to "small" emissions. Moderate insolation is assumed throughout the day and about 50% cloudiness is assumed at night. Thus wind speed and whether it is day or night will determine the category. Under very cloudy or very clear skies this scheme gives values about one-half a stability class lower than the Pasquill scheme. On an annual basis the Briggs scheme overestimates the frequency of A and B stability and underestimates the frequency of D and F stability, as determined by the STAR program for Edmonton International Airport.

1.3.5.2 Fluctuation Methods. Wind direction fluctuations over an hour as measured by an aerovane at 108 m have been used at Brookhaven National Laboratories (BNL) to identify "gustiness classes" (Singer and Smith 1953). The use of an instrument with a different sensitivity and chart speed would necessitate redefinition of the classes because a different portion of the eddy spectrum would be displayed. Also the records were obtained more than 100 m above the ground where the turbulence and wind speeds are quite different from those near the ground. Because of the anisotropy of boundary layer turbulence, a categorization of horizontal fluctuations may not be a suitable representation of vertical fluctuations. Nevertheless, wind direction fluctuations (azimuth angle) are relatively easy to measure and data are often available (for example, Islitzer and Slade 1968). Table 1.9 shows a composite scheme (Mitchell 1982, EPA 1986) which accounts

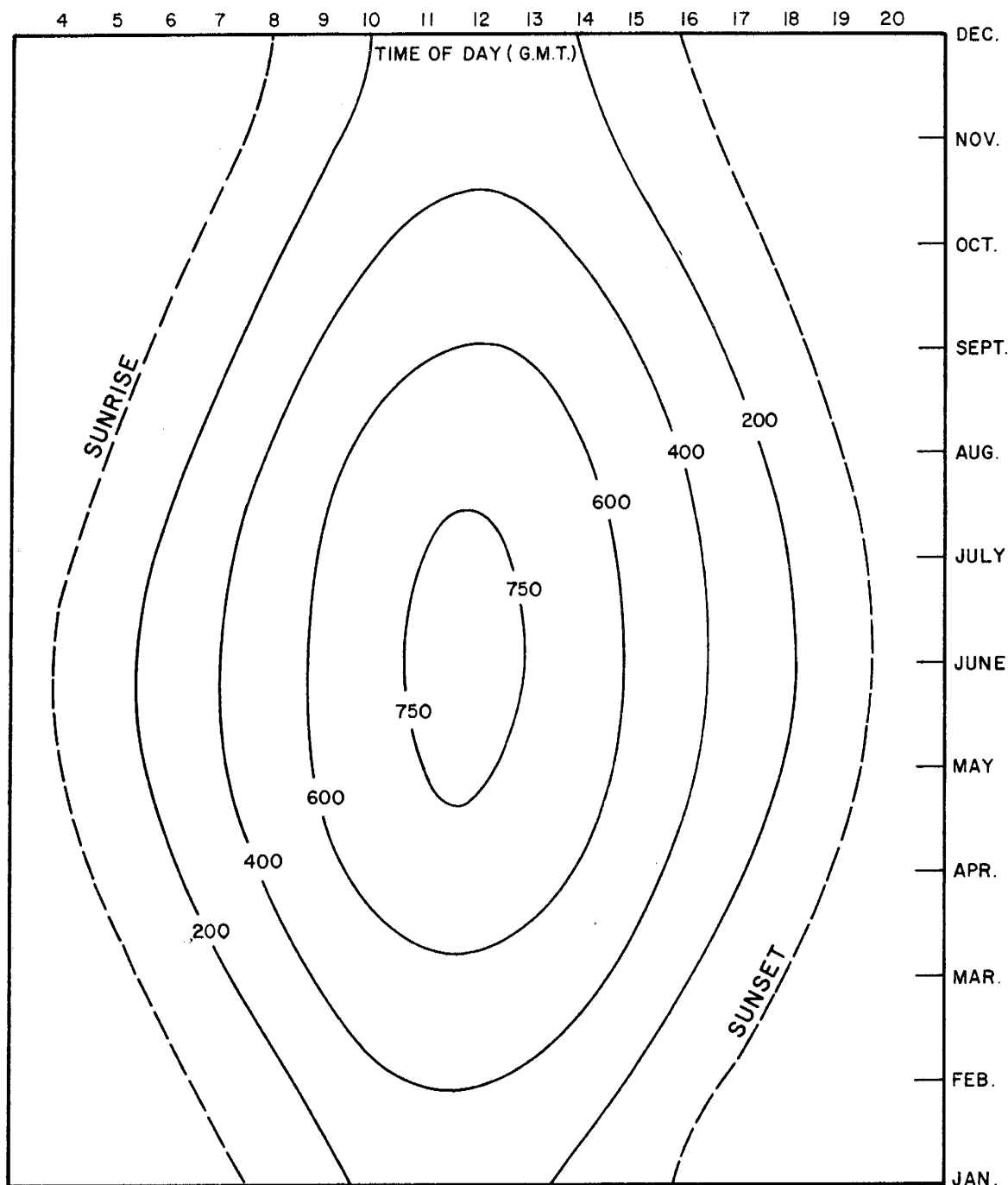


Figure 1.30 Solar radiation by month, hour and cloud cover. The incoming solar radiation in watts per square metre reaching the ground on a cloudless day in the U.K. is expressed in the main figure in terms of time of day and month. For cloudy conditions multiply this amount by the factor appropriate to the cloud amount in eighths.

Cloud Amount	Multiply I.S.R. by:
0	1.07
1	0.89
2	0.81
3	0.76
4	0.72
5	0.67
6	0.59
7	0.45
8	0.23

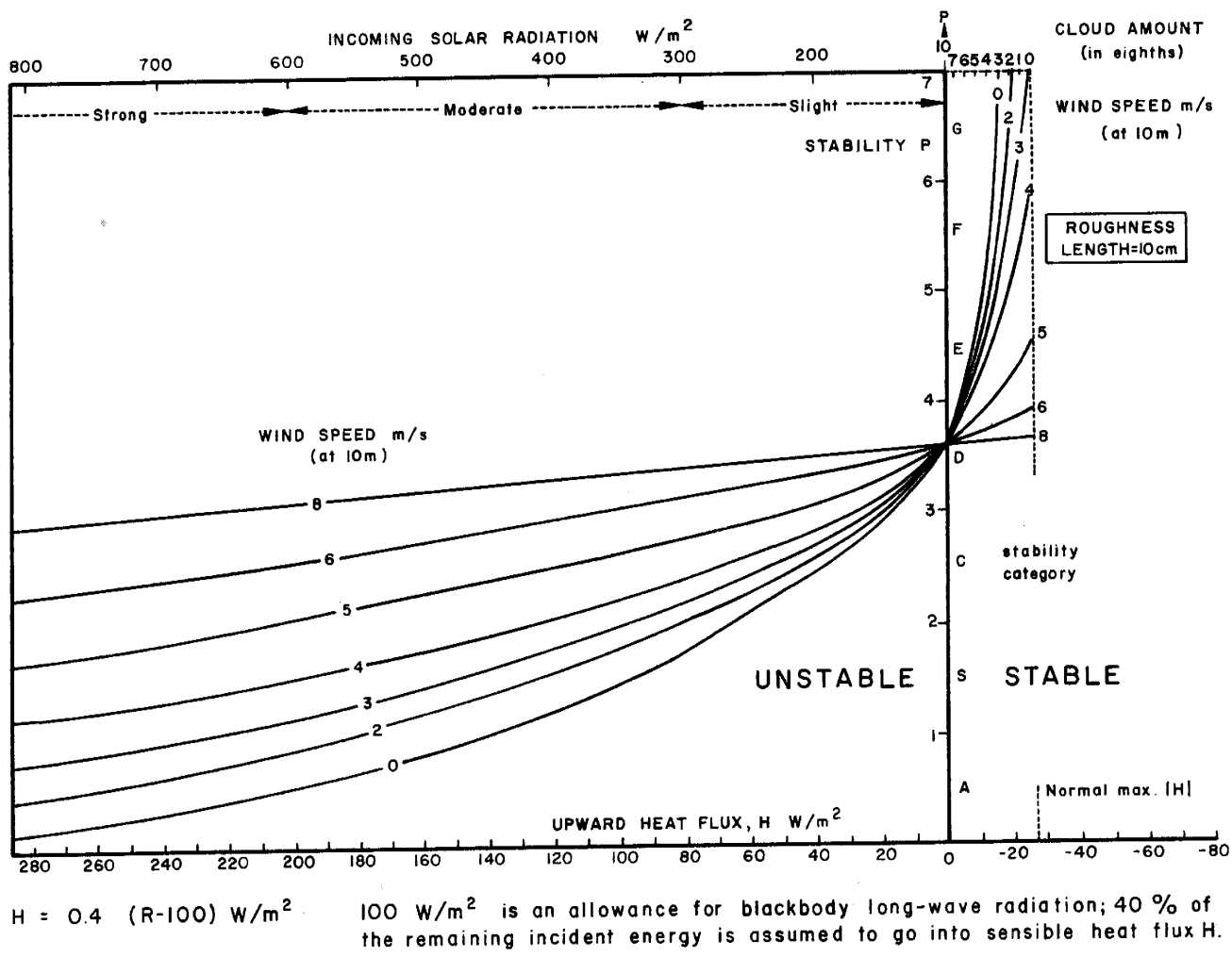


Figure 1.31 Stability number P from heat flux and wind speed. Stability categories A to G are prescribed in terms of upward heat flux (daytime) or cloud amount (nighttime) and surface wind speed for roughness length 10cm (Smith 1972).

Table 1.8 **Stability classification using net radiation and wind speed (after Williamson and Krenmeyer 1980).**

Net Radiation (W/m ²)	Wind Speed (m/s)									
	≥ 1.5	2	2.5	3	3.5	4	4.5	5	5.5	≥ 6
≥ 560	A	A	A	A	B	B	B	B	C	C
490	A	A	A	B	B	B	B	C	C	C
420	A	A	A	B	B	B	C	C	C	C
350	A	A	A	B	B	C	C	C	C	C
280	A	A	B	B	C	C	C	C	C	D
210	B	B	B	C	C	C	C	D	D	D
140	B	B	C	C	C	D	D	D	D	D
70	C	C	C	D	D	D	D	D	D	D
0	D	D	D	D	D	D	D	D	D	D
- 14	G	F	F	E	E	D	D	D	D	D
- 28	G	F	F	E	E	E	D	D	D	D
- 42	G	F	F	E	E	E	E	D	D	D
- 56	G	F	F	F	E	E	E	E	D	D
≥ 70	G	F	F	F	E	E	E	E	E	E

Table 1.9 Turbulence classification based on wind fluctuations (after EPA 1986, Mitchell and Timbre 1979).

Stability Description	Standard Deviation of the Horizontal Wind Direction Fluctuations (in degrees)	Standard Deviation of the Vertical Wind Direction Fluctuations (in degrees)
very unstable	Greater than 22.5°	Greater than 11.5°
moderately unstable	17.5° to 22.5°	10.0° to 11.5°
slightly unstable	12.5° to 17.5°	7.8° to 10.0°
neutral	7.5° to 12.5°	5.0° to 7.8°
slightly stable	3.8° to 7.5°	2.4° to 5.0°
moderately stable	2.1 to 3.8°	1.2 to 2.4°
very stable	< 2.1	< 1.2

Notes

- (1) These criteria are appropriate for steady-state conditions, a measurement height of 10 m, level terrain, and an aerodynamic surface roughness length of 15 cm. Care should be taken that the wind sensor is responsive enough for use in measuring wind direction fluctuations.
- (2) A surface roughness factor of $(z_0/15 \text{ cm})^{0.2}$, where z_0 is the average surface roughness in centimetres within a radius of 1 - 3 km of the source, may be applied to the tabulated values.
- (3) For nighttime hours with horizontal wind fluctuations indicated to be unstable, the following corrections should be applied:

If the indicated Stability Category is	and the Wind Speed at 10 m is (m/s)	then the Corrected Stability Category is
extremely unstable	< 2.4	very stable
	2.4 to 2.9	moderately stable
	2.9 to 3.6	slightly stable
	≥ 3.6	neutral
moderately unstable	< 2.4	moderately stable
	2.4 to 3.0	slightly stable
	≥ 3.0	neutral
slightly stable	< 2.4	slightly stable
	≥ 2.4	neutral

for the large standard deviations that often occur at night when the wind meanders (due to large quasi-two-dimensional eddies).

Both the azimuth and elevation angle fluctuations of a bivane have also been used to set up categories (Cramer in Gifford 1976). Bivane measurements, however, have not been common even during many detailed diffusion experiments.

1.3.5.3 Stability Index Estimation. The relationships between the fluxes of heat and momentum and the vertical profile of wind and temperature are now fairly well known. The relevant equations can be used to compute values of R_f , R_i or L from a few relatively simple measurements.

One practical scheme uses the wind speed at standard height and the temperature difference between this height and some lower height (Hanna et al. 1977, Wang et al. 1978, Wang 1981, Wilczak and Phillips 1984, Wratt 1987, Trombetti et al. 1986, Irwin and Binkowski 1981).

The link to the stability index is then made via the bulk Richardson number:

$$B = \frac{g z \Delta\theta}{\theta u^2}$$

where g is the acceleration due to gravity and z is the height of measurement of the wind u and the upper potential temperature θ in the potential temperature difference $\Delta\theta$. This is a more practical definition than the one used by Golder (1972). Because the factor gz/θ is nearly constant, sometimes the stability ratio $\Delta\theta/u^2$ is used. Bulk Richardson numbers for the Edmonton and Calgary tall towers are given in Table 1.10. Calgary exhibits a larger range than Edmonton.

Another alternative uses routine aviation weather observations and heat flux estimated from a detailed energy budget (van Ulden and Holtslag 1985, Schayes 1982, Holtslag and van Ulden 1983, Berkowicz and Prahm 1982, OME 1985, Alp et al. 1990, Ning and Yap 1986). Using estimates of the heat flux and roughness length, the profile equations are solved iteratively to determine L .

A third alternative defines a new stability length, L_n which is closely related to L , but uses more easily measured variables such as the net radiation and the wind speed (Briggs 1982).

Table 1.10 Bulk Richardson numbers measured on 91-m towers in Calgary and Edmonton (Choukalos 1985).

HOUR	SEASON							
	Winter		Spring		Summer		Autumn	
	Calgary	Edmonton	Calgary	Edmonton	Calgary	Edmonton	Calgary	Edmonton
1	2.94	0.46	1.52	0.58	1.46	0.76	1.75	0.50
2	2.46	0.52	1.39	0.38	1.97	0.64	1.64	0.45
3	2.14	0.85	1.18	0.59	1.83	0.50	1.59	0.79
4	1.75	0.91	1.33	0.53	1.28	0.74	1.63	0.67
5	2.54	0.36	1.87	0.44	1.32	0.52	1.83	0.85
6	3.03	0.63	1.33	0.53	1.50	0.44	1.17	0.60
7	2.27	0.75	1.48	0.44	0.66	0.38	1.83	0.79
8	2.45	1.10	0.76	0.28	0.29	0.07	1.46	0.69
9	1.93	0.47	0.12	0.19	1.17	0.14	0.90	0.41
10	2.02	0.44	0.68	0.00	1.78	0.34	0.36	0.32
11	1.22	0.49	1.00	0.28	1.30	0.36	0.25	0.21
12	0.12	0.07	1.04	0.31	1.37	0.37	1.22	-0.20
13	0.03	0.07	0.99	0.29	1.11	0.38	0.72	-0.25
14	0.82	0.03	0.84	0.28	0.96	0.21	0.99	-0.21
15	0.01	0.01	0.73	0.18	0.47	0.24	0.27	-0.16
16	0.20	0.14	0.27	0.17	0.40	0.25	0.12	-0.17
17	0.73	0.28	0.40	0.26	0.44	0.28	0.15	-0.11
18	1.55	0.42	0.29	0.14	0.25	0.14	0.30	0.32
19	1.74	0.53	0.04	0.05	0.12	0.07	0.92	0.58
20	2.21	0.37	0.68	0.08	0.04	0.11	1.31	0.60
21	2.50	0.55	0.68	0.53	0.26	0.46	1.41	0.61
22	1.66	0.35	0.64	0.38	0.62	1.03	1.73	0.52
23	2.54	0.88	0.75	0.31	0.55	0.86	1.66	0.44
24	2.68	0.43	1.14	0.44	1.02	0.77	1.85	0.42

1.3.5.4 Interrelationships and Comparisons. Because the different typing schemes all strive to represent the mix of mechanical and thermal turbulence, many workers have attempted to establish equivalencies as in Table 1.11 (Gifford 1976, Pasquill and Smith 1971). Observations from five sites (Golder 1972) showed that Turner's scheme (STAR) differed from Pasquill's in that Turner classes spread over two to four Pasquill classes. Golder's nomogram relating Pasquill class to Monin-Obukhov length and roughness can be fitted numerically (Liu and Durran 1977) as in Table 1.12.

The Pasquill class P and the Kazanski-Monin scale m (also called the planetary Richardson number) are related by (Smith 1979):

$$P = \frac{3.6}{1 - 0.53 (\mu/100) + 4.9 (\mu/100)^2 - 2.0 (\mu/100)^3}$$

at latitude 51° N for neutral and unstable cases. This has been extended to stable conditions and a variety of roughness lengths (Tagliazucca and Nanni 1983) as shown in Table 1.13. This information fits the relationship (Sutherland et al. 1986)

$$P = 3.6 \exp (\mu/a)$$

where $a = 10 \ln (7.5/z_0)$ and z_0 is the roughness length in m.

While there may be a correspondence between median values in each class, the frequency distributions from different schemes are often found to differ markedly (Smith 1974, Skaggs and Robinson 1976, Fulle 1976, DeMarrais 1978, Sedefian and Bennett 1980, Reiquam 1980, Kretzschmar and Mertens 1979, Atwater and Londergan 1985, Gildart and Chang 1983, Draxler 1987) more often than they are found to be similar (Bowen et al. 1983, Mitchell 1982). Terrain factors are often cited as the cause of these differences (Rao and Roffman 1977, Sandusky 1985, Lalas et al. 1979).

Some workers have tried to use temperature gradient alone as a stability parameter (US NRC 1972, Carpenter et al. 1971, Portelli 1976). The results are sensitive to the height interval (Pendergast and Crawford 1974, Courtney 1977, Reynolds 1978) because the gradients tend to become dry adiabatic with increasing altitude. Empirically there is no relation between temperature gradient alone and turbulence (Luna and Church

Table 1.11 Relations among turbulence typing methods (Gifford 1976, Pasquill and Smith 1971).

Stability Description	Pasquill	Turner	BNL	σ_A (deg)	Ri (at 2 m)	L (m) ($z_0 = 1$ cm)
Very unstable	A	1	B ₂	25	-1.0 to -0.7	-2 to -3
Moderately unstable	B	2	B ₁	20	-0.5 to -0.4	-4 to -5
Slightly unstable	C	3	B ₁	15	-0.17 to -0.13	-12 to -15
Neutral	D	4	C	10	0	∞
Slightly stable	E	6		5	0.03 to 0.05	35 to 75
Moderately stable	F	7	D	2.5	0.05 to 0.11	8 to 35

Table 1.12 Relation between Pasquill class, the reciprocal of the Monin-Obukhov length ($1/L$) and roughness length (Liu and Durran 1977).

Pasquill Category	$1/L = a z_0^b$		z_0 (m)		
	a	b	0.01	1	0.1
				$1/L$ (m ⁻¹)	
A	-0.1135	-0.1025	-0.18		-0.14
B	-0.0385	-0.1710	-0.08		-0.06
C	-0.0081	-0.3045	-0.03		-0.02
D	0	-0.5030	0		0
E	0.0081	-0.3045	0.03		0.02
F	0.0385	-0.1710	0.08		0.06

Table 1.13 Values of Kazanski-Monin parameter corresponding to Pasquill classes (after Sutherland et al. 1986).

Pasquill Class	Kazanski - Monin Parameter			
	Roughness Length (m)			
	0.001	0.01	0.10	1.00
A	< -85.9	< -75.6	< -60.4	< -44.0
B	-55.0 to 85.9	-45.3 to -75.6	-31.6 to -60.4	-13.7 to -44.0
C	-31.5 to -55	-20.6 to 45.3	-12.4 to -31.6	-5.5 to -13.7
D	+15.1 to -31.6	+8.93 to -20.6	+4.81 to -12.4	+3.44 to -5.5
E	+33 to +15.1	+22.09 to +8.93	+13.7 to 4.81	+6.87 to 3.44
F	> 33.0	> 22.0	> 13.7	> 6.87

1972, Pendergast and Crawford 1974, Gifford 1976) except perhaps in stable conditions (Weber et al. 1977). In principle, such schemes are not expected to work because the wind shear component is missing.

In northeastern Alberta MEP (1976) found little correlation between various parameters. Near Edson, Davies and Thompson (1983) reported that on-site wind fluctuations gave higher frequencies of neutral and unstable classes than did STAR data for the Edson airport. Using a heat flux/profile method Alp et al. (1990) also found higher frequencies of unstable classes at Calgary compared to STAR.

1.4 WEATHER SYSTEMS

The atmosphere is a continuum in space and time. Like any turbulent fluid it has a wide variety of perturbations on different time and space scales. Certain of these perturbations repeat themselves again and again with sufficient similarity to permit them to be regarded as members of a discrete family with fairly well defined characteristics. Each system lies within a certain size range and has its own thermal and circulation patterns. Such organized weather systems may be classified by their extent in time and space as in Figure 1.32 (Orlanski 1975).

The three-cell system model of the general circulation is well known. Alberta lies in the zone of the mid-latitude westerlies. Strong rotational effects, large temperature contrasts and the preponderance of land masses with irregular relief (in the northern hemisphere) lead to the formation of mobile pressure systems near the earth's surface. These cyclones and anticyclones disrupt the westerly flow and as a result wind roses at weather stations often do not show a prevailing westerly. At higher altitudes, however, strong westerly components are present and account for the more or less west to east movement of the migratory pressure systems.

1.4.1 Airmass Systems

An *airmass* is formed when a large portion of the atmosphere remains over an extensive area of uniform surface features long enough for the air to take on certain characteristic values of temperature, moisture and stability. In Canada it has become conventional to identify four major airmasses (Anderson et al. 1955, Penner 1955): maritime tropical, maritime polar, maritime Arctic, continental Arctic. The characteristics

TYPICAL TIME		1 MONTH	1 DAY	1 HOUR	1 MINUTE	1 SECOND	
TYPICAL DISTANCE							
10,000 km 2,000 km 200 km 20 km 2 km 200 m 20 m		STANDING WAVES	ULTRA-LONG WAVES				MACRO α SCALE
			BAROCLINIC WAVES				MACRO β SCALE
			FRONTS & HURRICANES				MESO α SCALE
				NOCTURNAL LOW LEVEL JET SQUALL LINES INERTIAL WAVES CLOUD CLUSTERS MTN. & LAKE DISTURBANCES			MESO β SCALE
				THUNDERSTORMS CLEAR AIR TURBULENCE URBAN EFFECTS			MESO γ SCALE
				TORNADOES DEEP CONVECTION SHORT GRAVITY WAVES			MICRO α SCALE
				DUST DEVILS THERMALS WAKES			MICRO β SCALE
					PLUMES ROUGHNESS TURBULENCE		MICRO γ SCALE

Figure 1.32 Scale definitions and different processes with characteristic time and horizontal scales (after Orlanski 1975).

Table 1.14 Airmasses and fronts in the conventional Canadian classification.

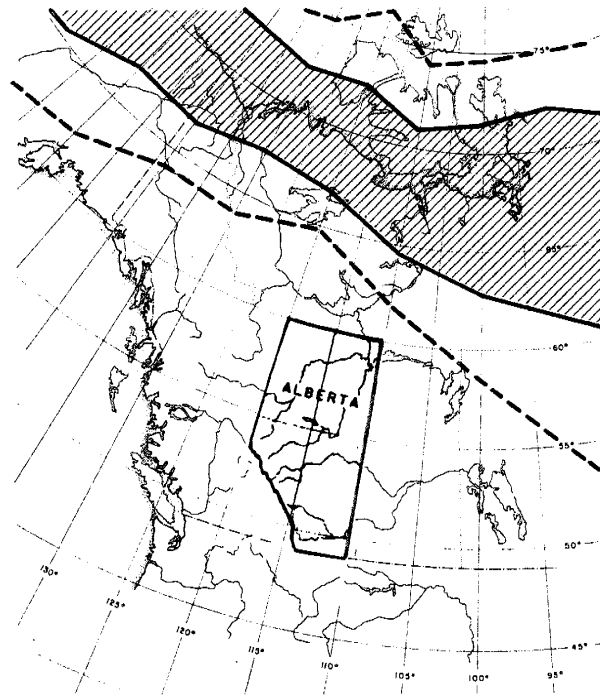
Property	Airmass			
	Continental Arctic	Maritime Arctic	Maritime Polar	Maritime Tropical
Source Region	Arctic ice cap frozen ocean snow covered land	Arctic Ocean	Mid-latitude oceans	Subtropical oceans
Typical Winter Temperature (°C)	-25	-4	6	15
Typical Summer Temperature (°C)	N/A	2	14	18
Front Separating from Warmer Airmass	Arctic	Maritime	Polar	---

of these airmasses are summarized in Table 1.14. As airmasses move in response to changes in the large-scale circulation they pass over different types of surfaces with subsequent modifications of the airmass properties until ultimately the airmass loses its identity. Thus some meteorologists may identify an airmass as "modified," "warm" or "cold." The four-airmass convention is not used universally. In the U.S. and elsewhere other naming conventions are adopted.

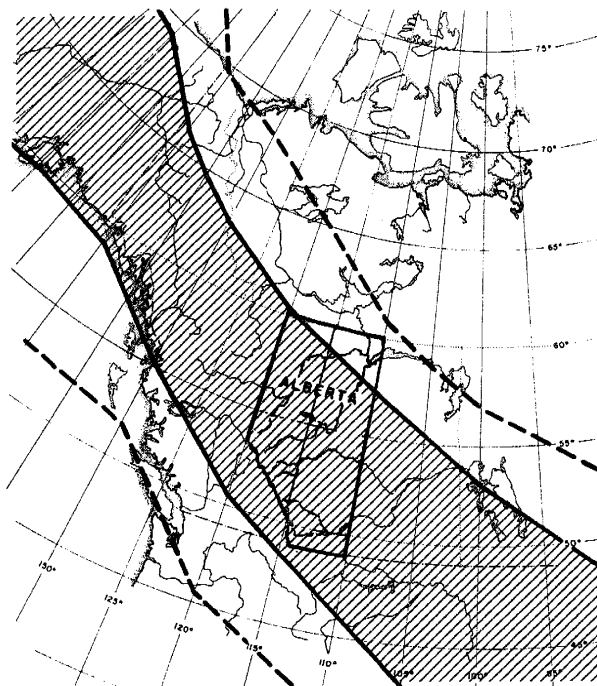
The mean sea level pressure patterns for the northern hemisphere reveal the combined effects of large-scale circulation and airmass type. In the cold months of the year, Alberta weather may be affected by either of two semipermanent pressure systems. The Aleutian low dominates the Pacific Ocean off the coast of British Columbia and periodically "sheds" cyclones that bring storms to Alberta. The polar continental high dominates the Arctic region. It is a shallow formation and its appearance is to be interpreted as a statistical average rather than as an omnipresence (Petterssen 1956). Its relatively deep, intense surface-based inversions and light surface winds create a high pollution potential in the Arctic, at least with respect to low-level sources. In the warm months, the North Pacific high dominates the whole area off the coast of British Columbia. The subsidence inversions associated with this large, deep subtropical anticyclone are infamous in connection with California's air pollution problems. The polar continental high disappears in the warmer months, but there is a shallow anticyclone over the Arctic ice fields (Petterssen 1956).

The variability of the weather at Alberta's latitude is due to the boundaries between airmasses, known as *fronts*, the cyclones that form along these fronts, and the migratory anticyclones that are interspersed in the various sequences of events. In Canada it is conventional to identify three fronts separating the four airmasses: The polar front separating the maritime tropical from maritime polar air, the maritime (Arctic) front separating maritime polar air from maritime Arctic air, and the Arctic front separating maritime Arctic from continental Arctic air. The polar front in winter has an average position at about 25°N on the surface and seldom penetrates much beyond 45°N. The maritime front has an average surface position at about 40°N over the oceans and south of the snow line over the continent (Anderson et al. 1955).

In the winter the Arctic front lies across Alberta for much of the time while in spring and autumn it retreats to the northern border of the province. In summer it



July



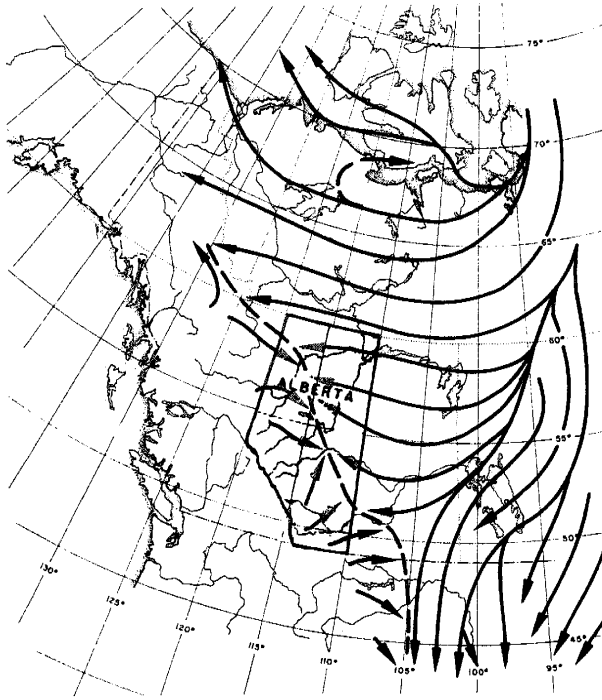
January

Figure 1.33 Quartiles (boundaries of hatched area) and deciles (dashed lines) of the 85-kPa Arctic front location for period 1961 - 1965 (Barry 1967).

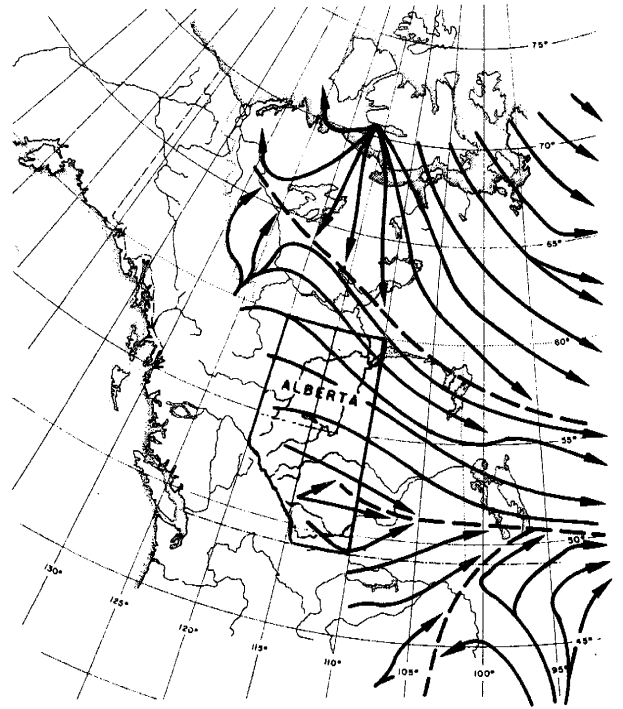
generally lies along the Arctic coastline (Barry 1967). Quartile and decile positions for winter and summer are given in Figure 1.33. Streamlines of the surface resultant wind can also be used to define the mean positions of airmasses and fronts (Bryson 1966). Since mean streamlines are also mean trajectories, the area occupied by a given airmass is also an area of streamlines leading back to the source region. Zones of streamline convergence or confluence then represent fronts (Figure 1.34). Airmass frequencies in July obtained by following airmasses back to their entry points and classifying them as Arctic, Pacific, United States and Atlantic (Bryson 1966) are shown in Figure 1.35. Alberta is dominated by Pacific air about 80% of the time, with about a 10% occurrence of Arctic air and about a 5% occurrence of United States air. Most of this United States air is ultimately of Pacific origin. The region of equal frequency between the Arctic and Pacific air can be regarded as the modal position of the front separating them.

When a front is *stationary* the winds blow parallel to the boundary and there will be higher pressures both in the cold air and the warm air. The frontal inversion associated with this situation may result in a lengthy period of plume trapping (see section 2.7).

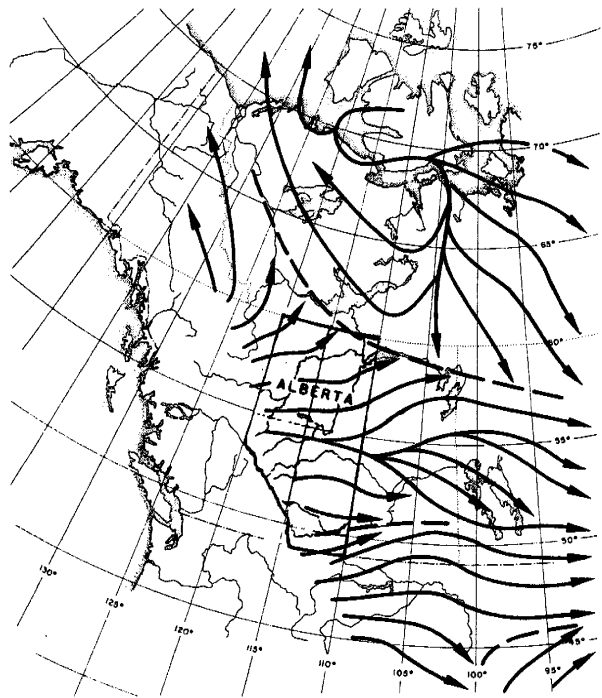
For reasons not fully understood a front may develop a "kink" which is associated with the formation of a cyclone. The westernmost section of the boundary is designated "cold front" because cold air is advancing and undercutting warm air. The eastern section of the boundary is designated "warm front" because the warm air is advancing and overrunning the cold air. The cold front has a steep slope, generally between 1/150 to 1/50, which causes vigorous uplifting and, like the cyclonic centre itself, is often associated with extensive cloud and precipitation. The winds behind a cold front are generally brisk and the front itself moves fairly rapidly. The warm front, on the other hand, has a shallow slope of 1/100 to 1/300. The air behind the warm front becomes stabilized through contact with the relatively cool ground over which it moves. Winds in the warm sector are light and the warm front moves rather slowly. As a result the presence of a warm front can present a serious pollution potential, at least for short time periods. In hilly country where pockets of cold air may remain in valleys after the warm front has passed, the frontal inversion will inhibit vertical dispersion. The surface winds associated with a warm front tend to be from the southeast quadrant, while those associated with a cold front tend to be from the northwest quadrant. Mathews et al. (1984) documented some of the characteristics of Arctic frontal passages near Calgary.



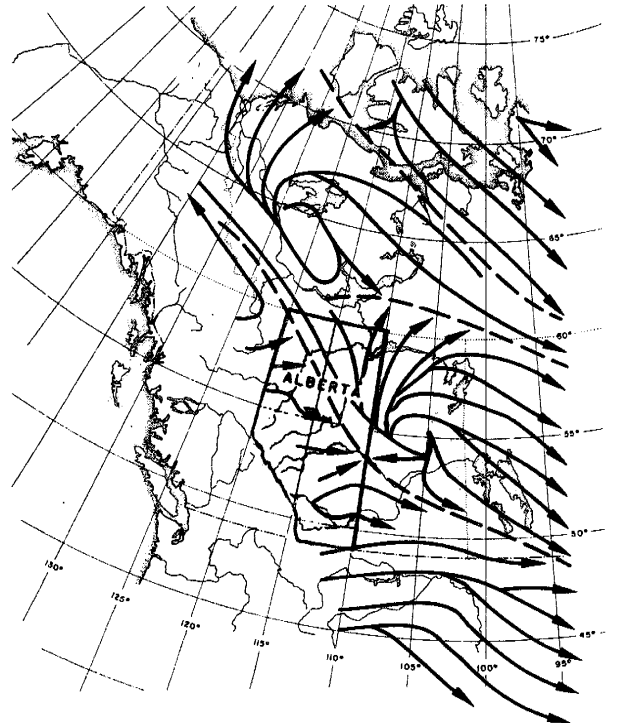
April



July

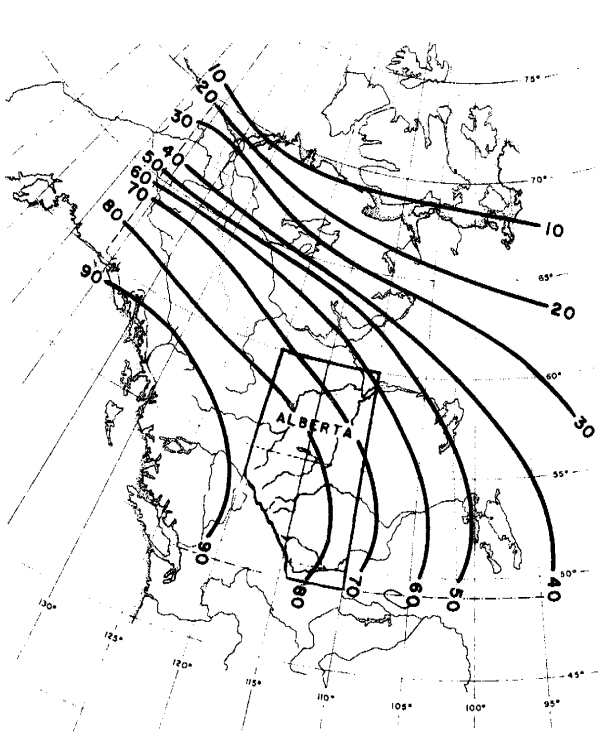


October

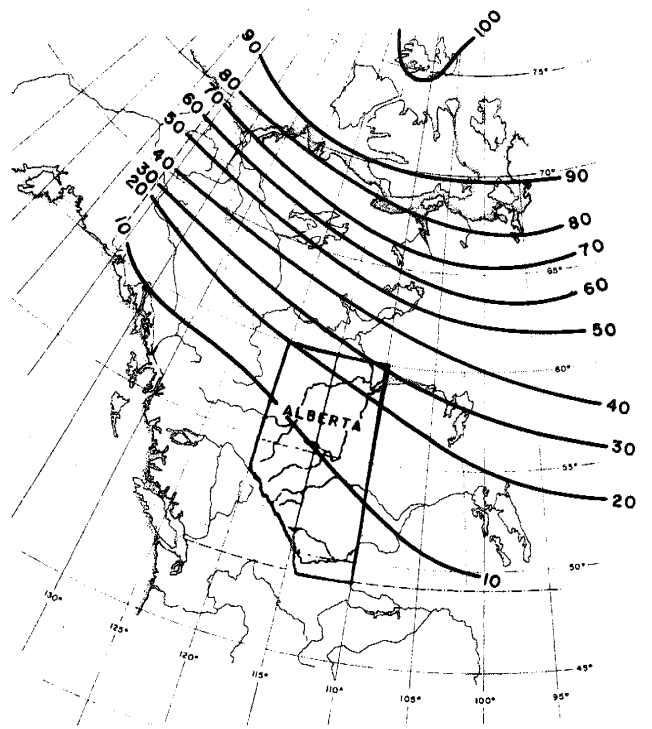


January

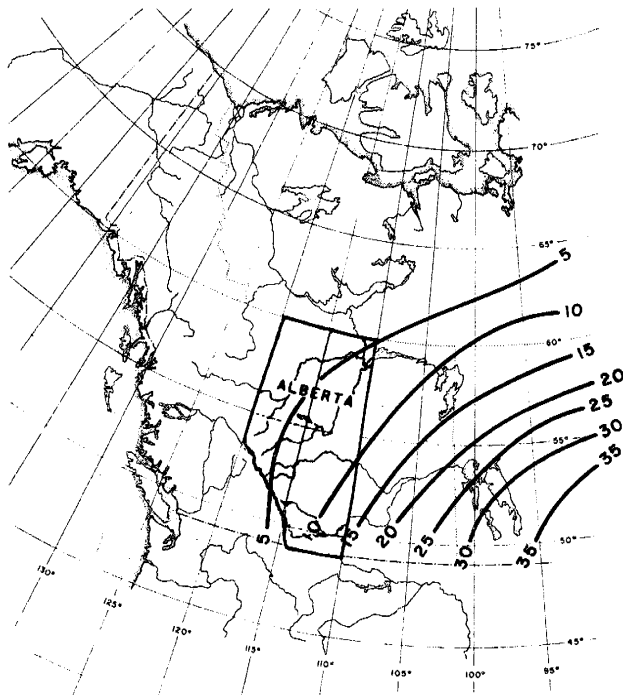
Figure 1.34 Streamlines of the surface resultant wind and zones of confluence (Bryson 1966).



Pacific air



Arctic air



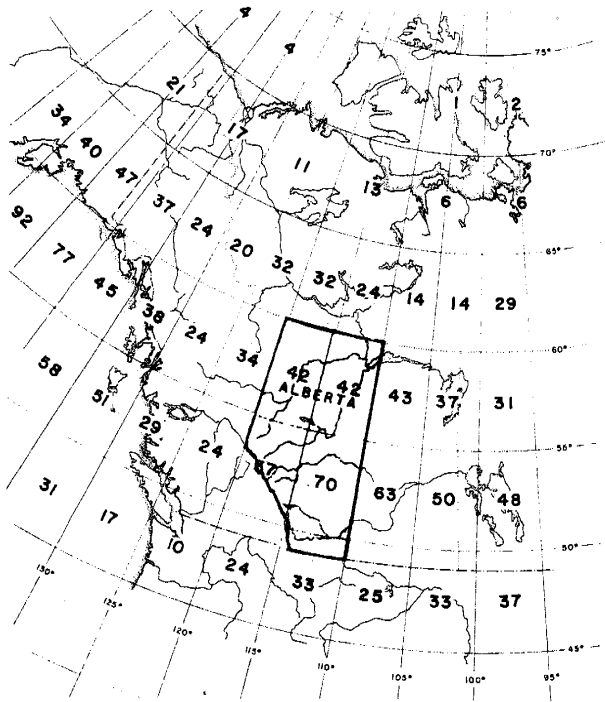
United States air

Figure 1.35 Isopleths of the percentage frequency of occurrence of different airmasses defined by their direction of entry (Bryson 1966).

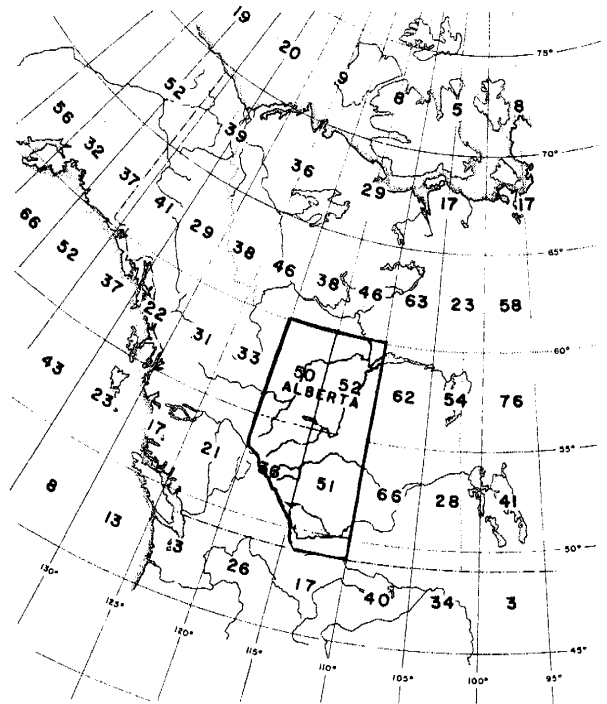
The number of days with low pressure centres in a 40-year period using cells 5° latitude in length and 5° longitude in width (Klein 1957) is given in Figure 1.36. Isopleths of the number of different low pressure centres in a 20-year period are shown in Figure 1.37. Alberta lies near a local maximum in every season. The main motions of low pressure systems in the northern hemisphere are indicated in Figure 1.38.

Most migratory anticyclones are offshoots of the semipermanent centres. "Warm" anticyclones are spawned by the Pacific high and tend to move very slowly eastward. These systems are characterized by light winds through a deep layer and widespread poor dilution (Holzworth 1969). They are accompanied by a ridge of high pressure aloft and sustained subsidence (Niemeyer 1960). In the United States most episodes of undesirable air quality (and some of the historical air pollution disasters) have been found to occur with these *stagnating anticyclones* (Miller and Niemeyer 1963, Holzworth 1962). However, as the principal anticyclone tracks show (Figure 1.39), these warm highs rarely affect Alberta, except perhaps in the summer.

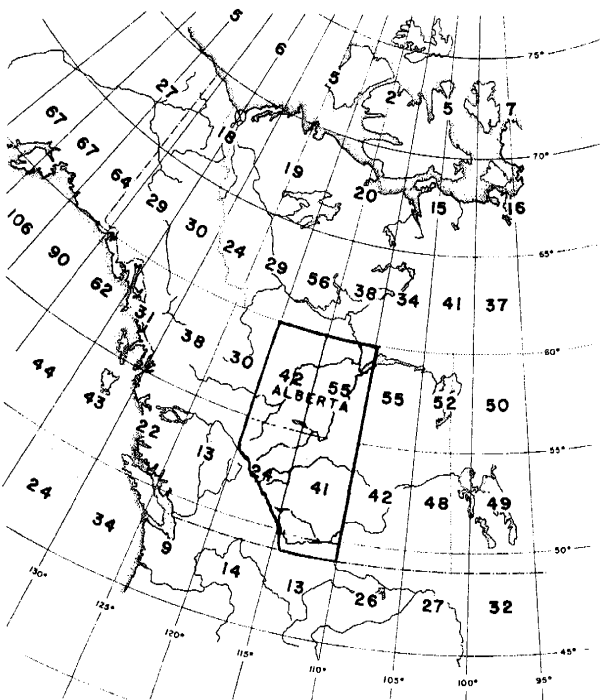
"Cold" anticyclones have their origins in the polar continental high of the northwestern Arctic. They tend to move southeastward across the Canadian prairies into the United States (Figure 1.39). They are generally located between migrating cyclones and are rarely stationary long enough to cause much of a pollution potential (Shaw and Munn 1971). In addition, they undergo heating from the relatively warm surfaces over which they move and are often (conditionally) unstable in their lowest layers. The atmosphere is relatively cold through a deep layer and the rapid transport of cold air aloft into the high tends to maintain a comparatively strong lapse rate and enhance vertical mixing (Holzworth 1969). Since winds normally increase with height over cold anticyclones, vertical mixing also enhances horizontal mixing by the downward transport of momentum (Holzworth 1969). As the cold anticyclone moves southward it becomes transformed and ultimately merges with the subtropical Atlantic high, sometimes creating a pollution potential in the eastern United States. The number of days with highs in a 40-year period (Klein 1957) is given in Figure 1.40. Isopleths of the number of different highs in a 20-year period are shown in Figure 1.41.



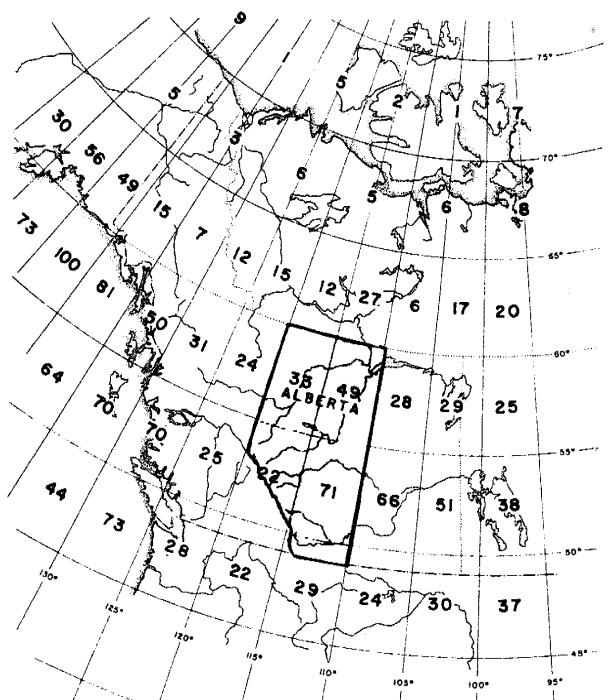
April



July



October



January

Figure 1.36 Number of days with lows in period 1899 - 1938 (Klein 1957).

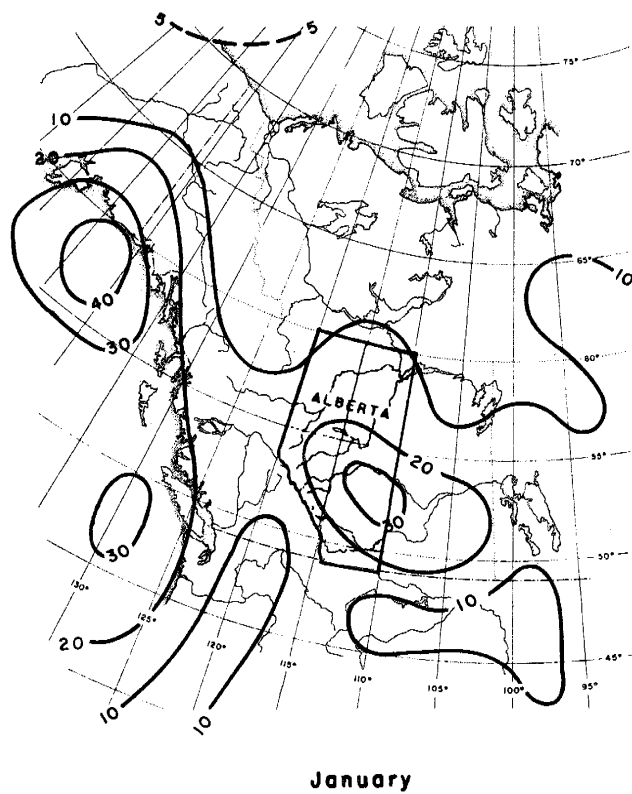
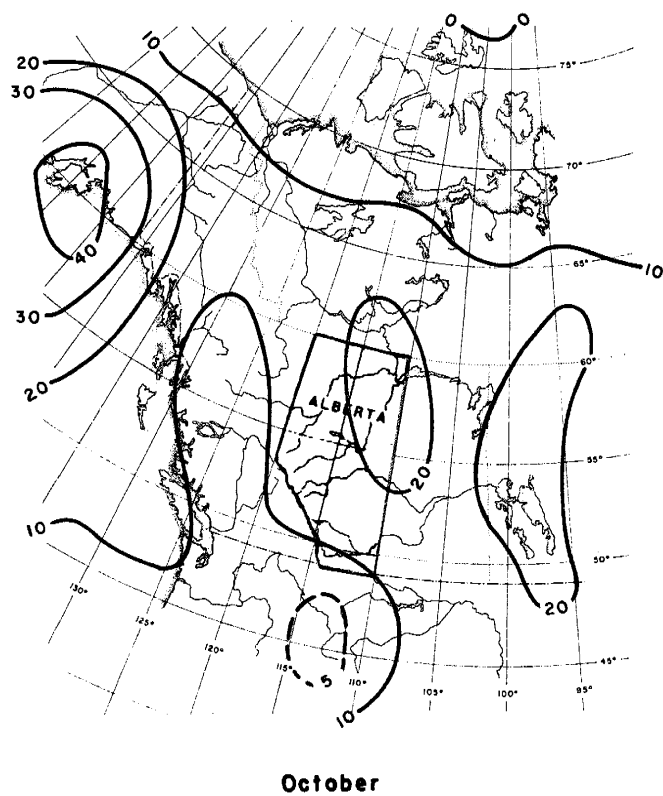
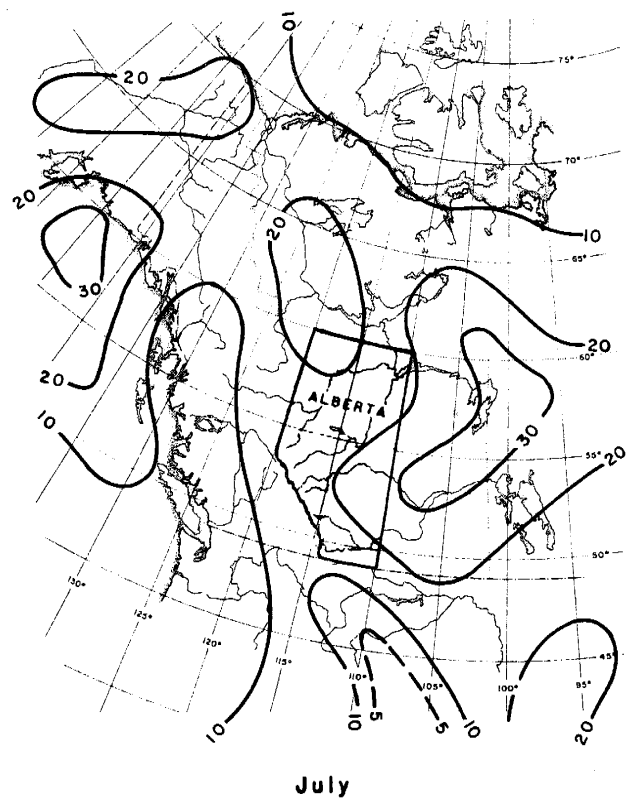
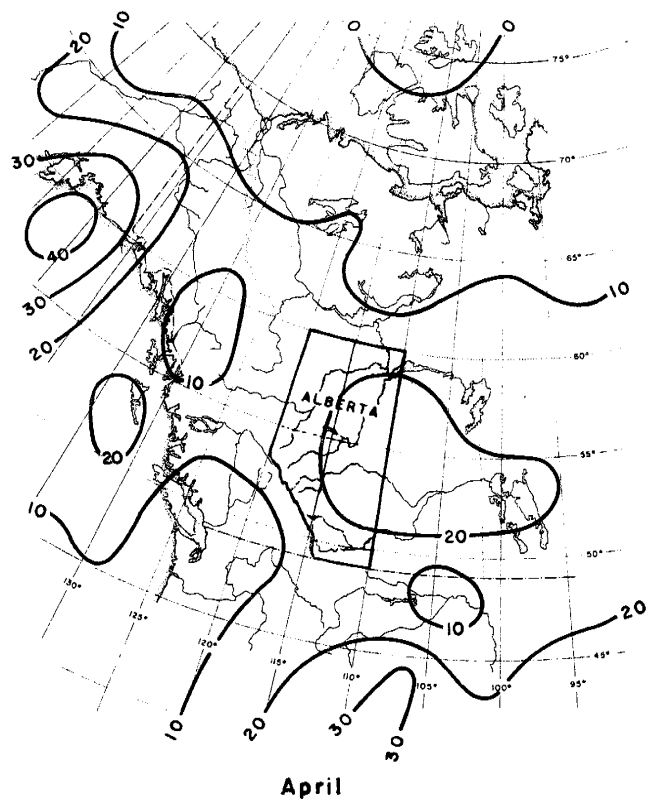
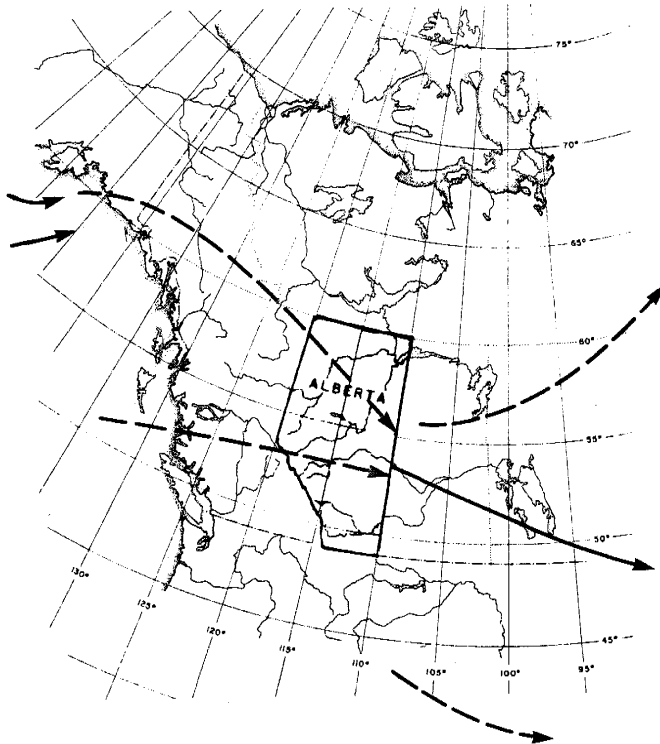
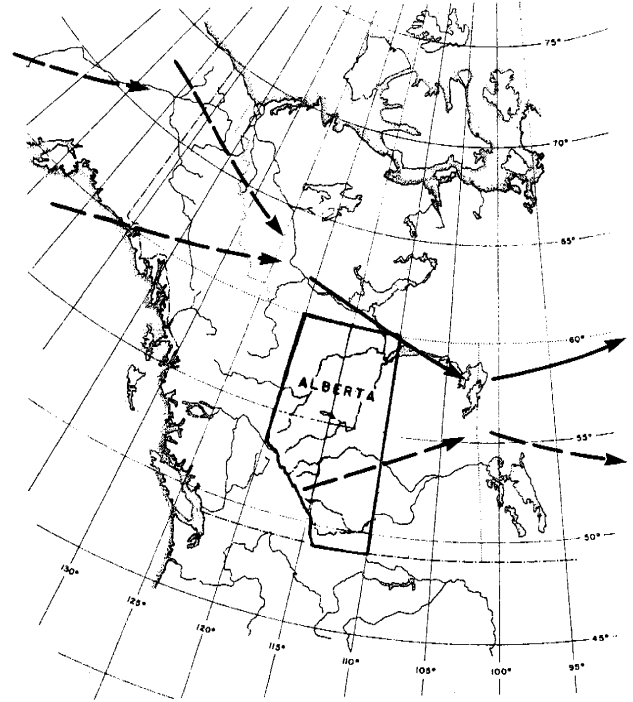


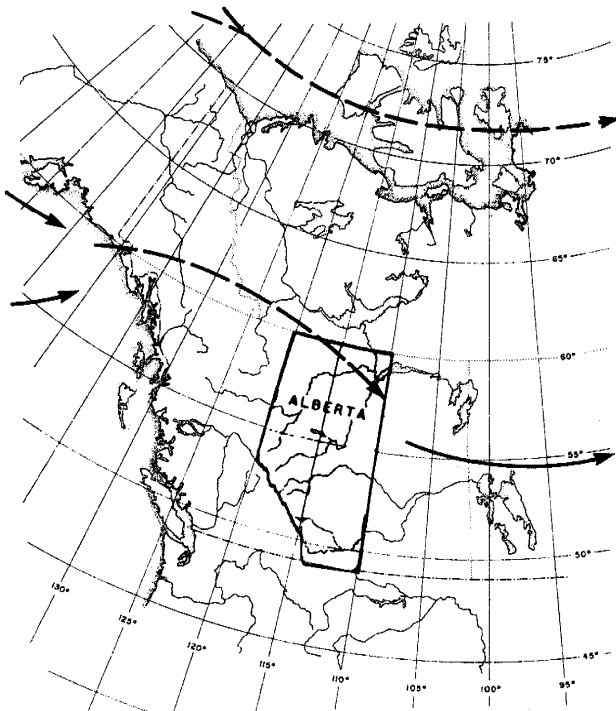
Figure 1.37 Isopleths of the number of different lows in period 1909 - 1914 and 1924 - 1937 (Klein 1957).



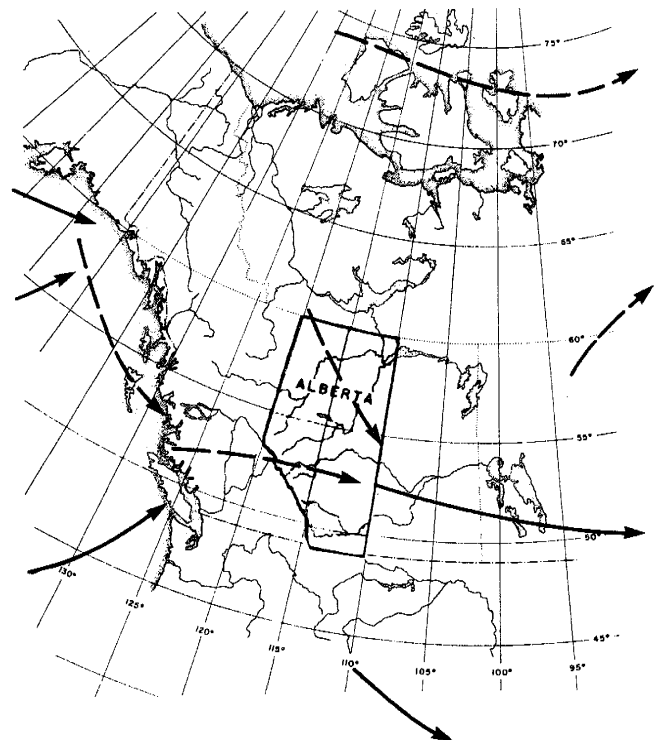
April



July

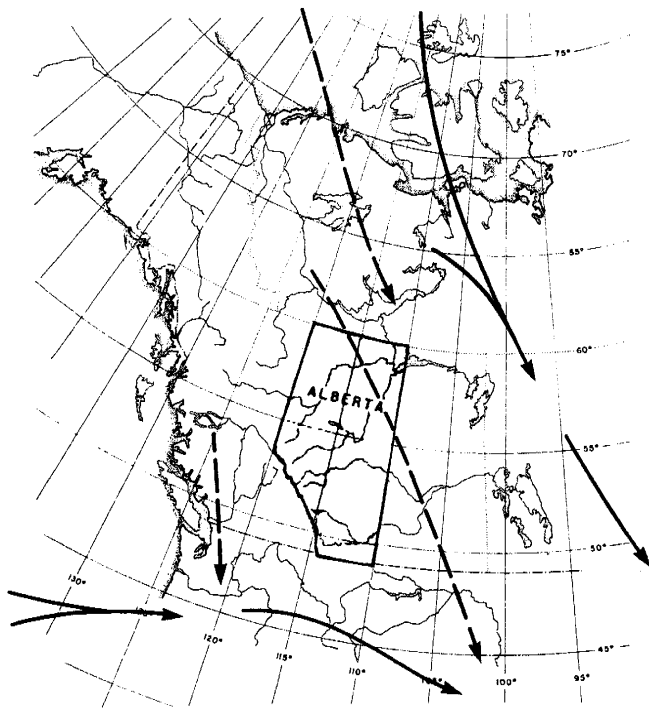


October

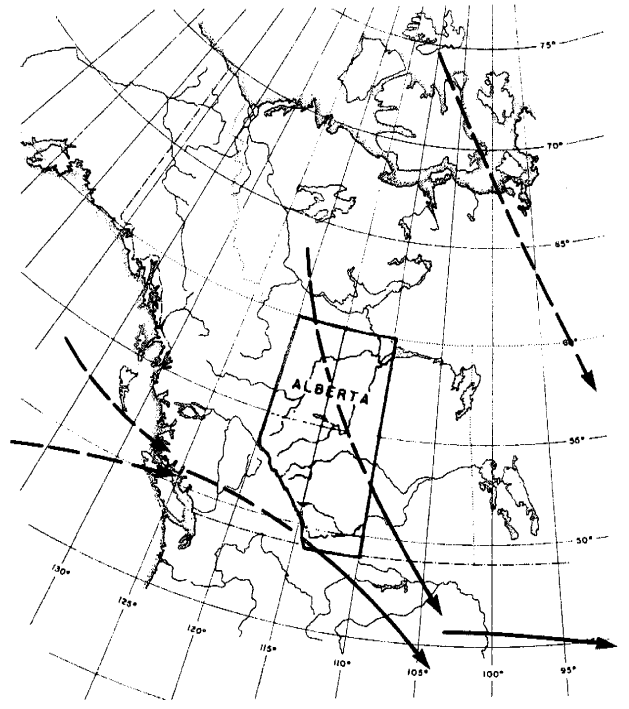


January

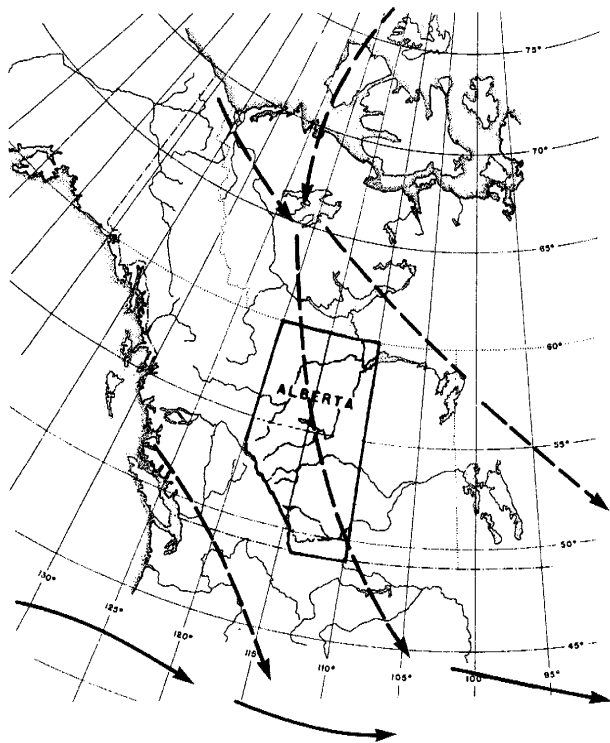
Figure 1.38 Principal tracks of lows in and around Alberta (after Klein 1957).



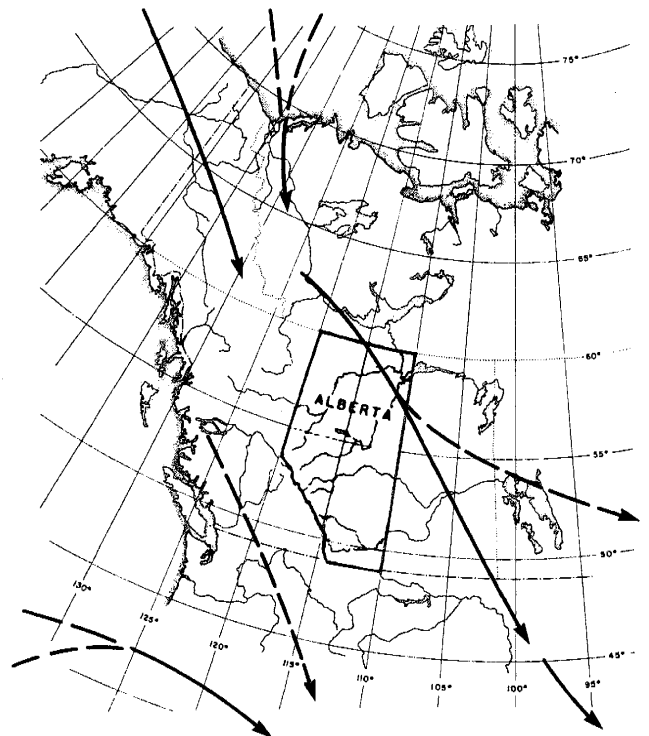
April



July

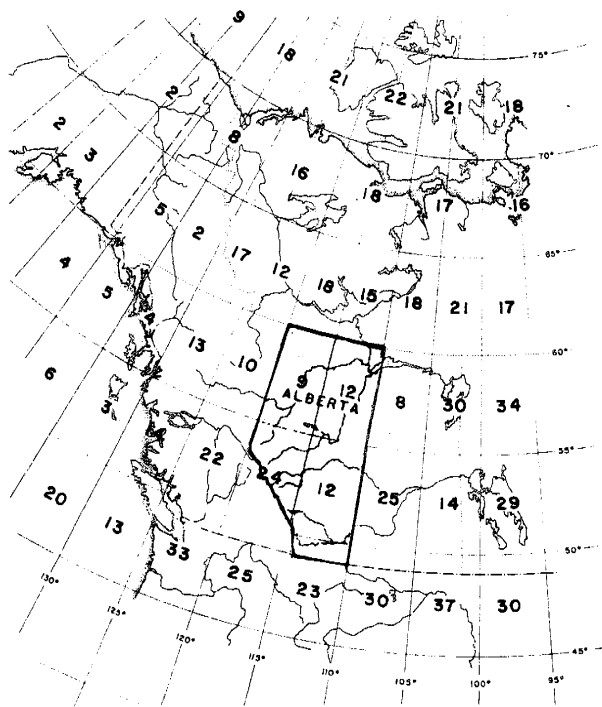


October

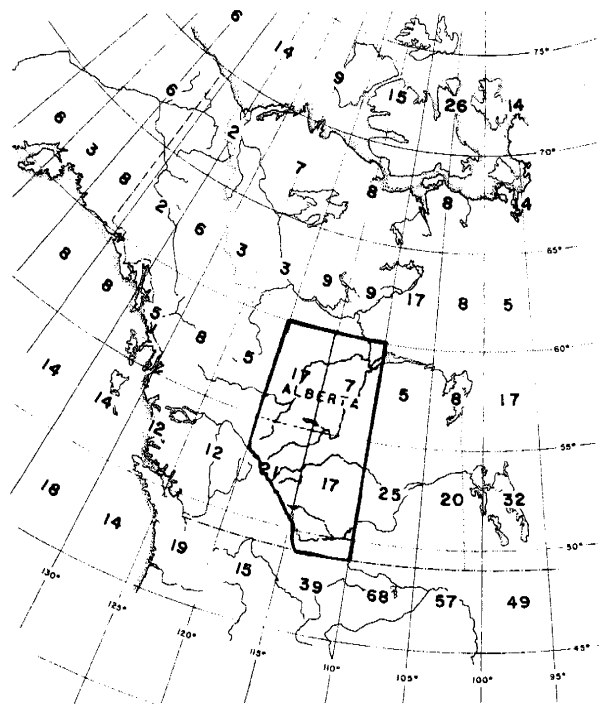


January

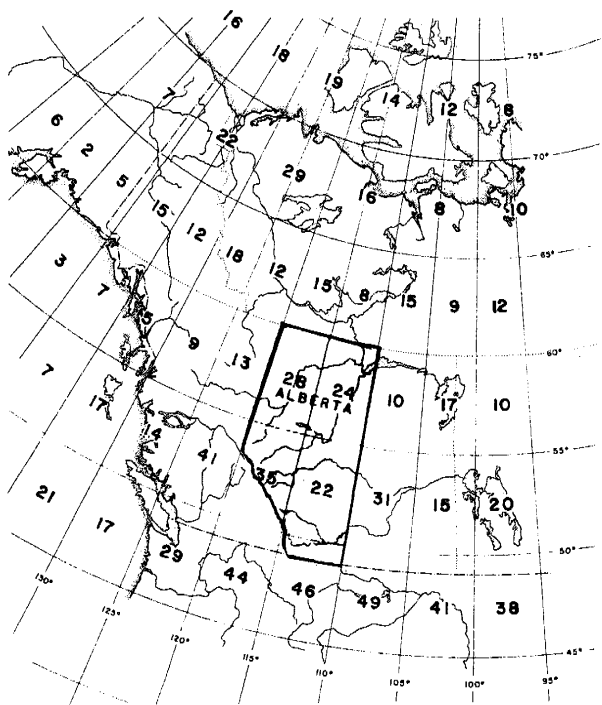
Figure 1.39 Principal tracks of highs in and around Alberta (after Klein 1957).



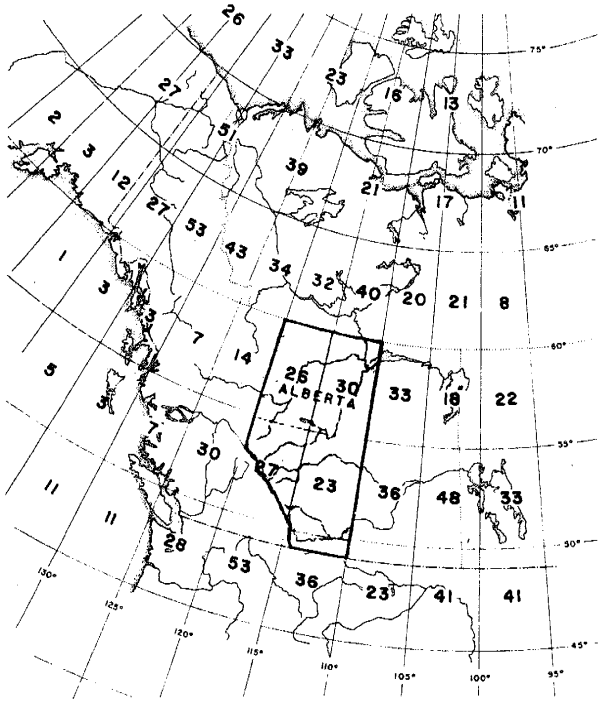
April



July

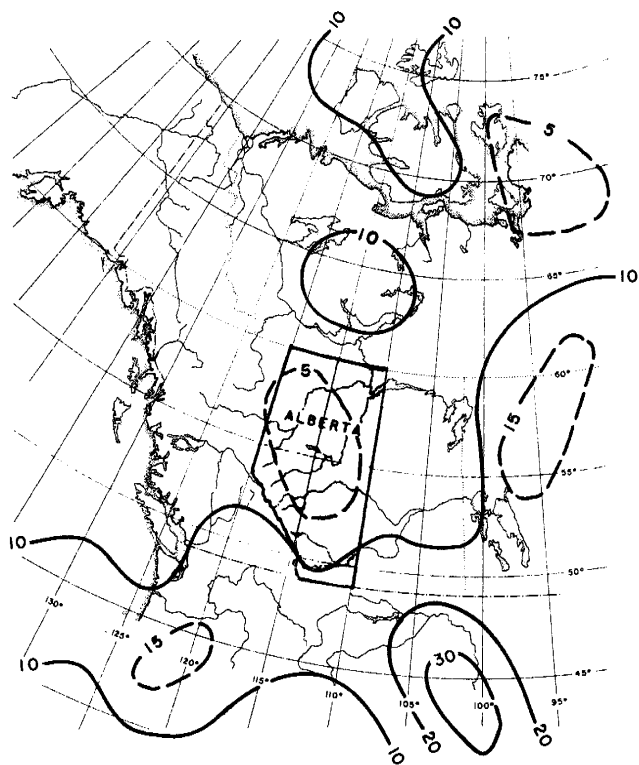


October

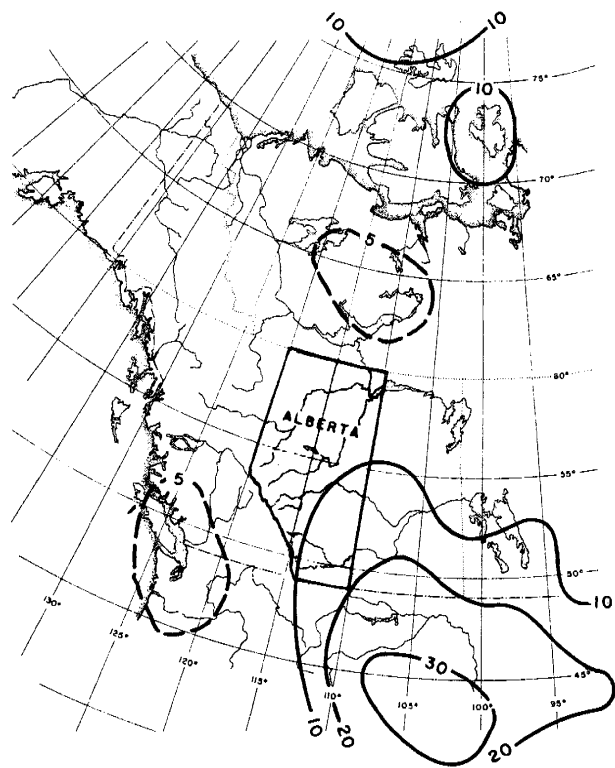


January

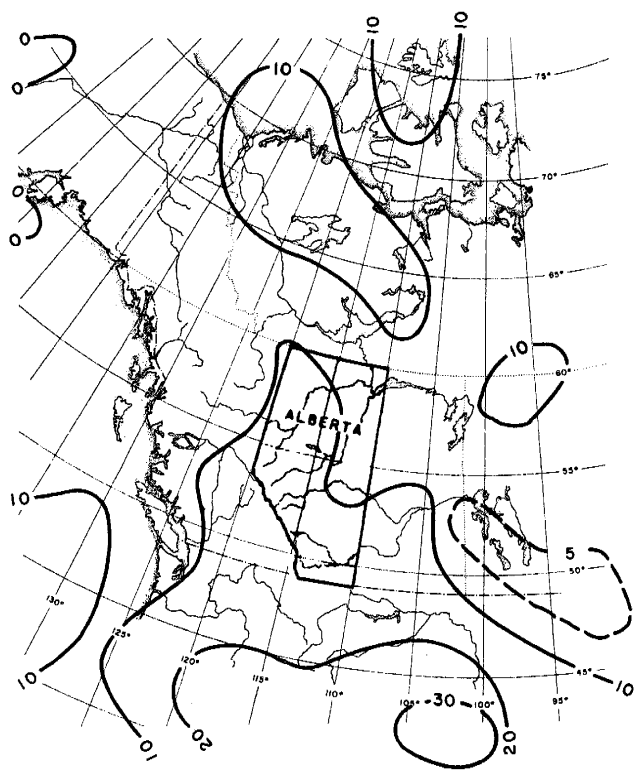
Figure 1.40 Number of days with highs in period 1899 - 1938 (Klein 1957).



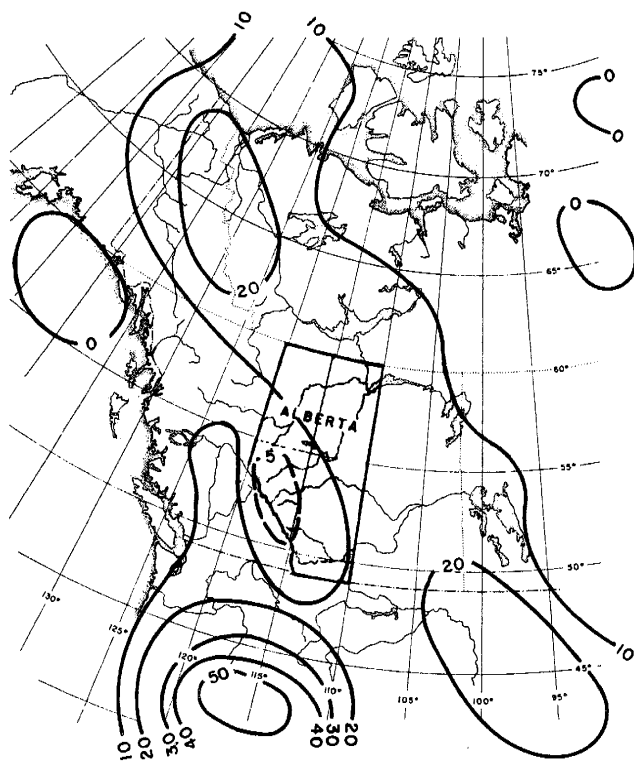
April



July



October



January

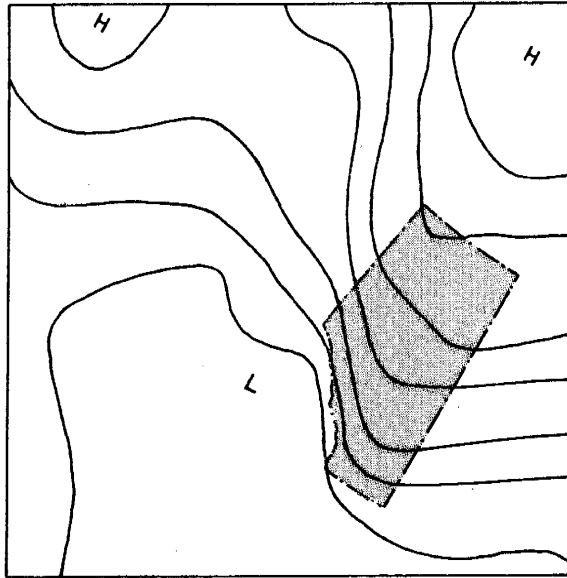
Figure 1.41 Isopleths of the number of different highs in period 1909 - 1914 and 1924 - 1937 (Klein 1957).

The pressure patterns on surface weather maps can be abstracted into a number of prototypes that describe or represent situations that occur frequently or which have important influences on some specific element or application. This process of pattern recognition or weather map typing has, until recently, been subjective and as often as not, purely subconscious in the mind of the weather forecaster. The advent of computers made possible a number of completely objective synoptic typing systems. One such scheme uses linear correlation between each map and all others in the data set. A correlation coefficient threshold value is chosen and a search is made through the correlation matrix for the map with the most correlations larger than the threshold. This map, and all others correlated with it at or above the threshold, are designated as "Type 1". These maps are then removed and the procedure is repeated to obtain "Type 2" etc. until no more maps can be assigned to a type. The remaining maps are designated as "uncorrelated."

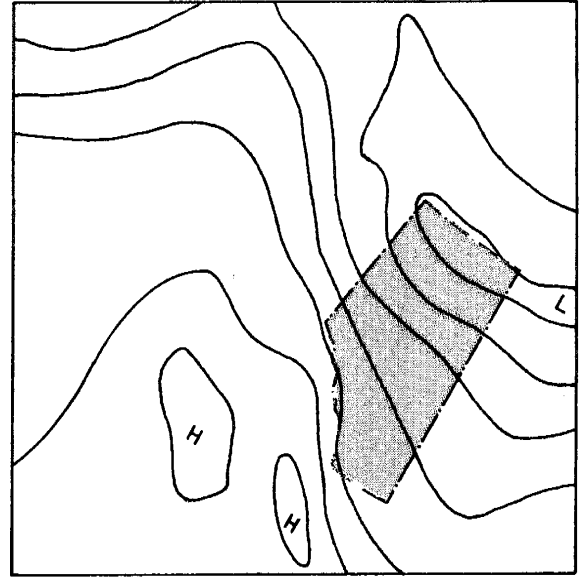
Kociuba (1974) developed 33 surface map types (plus 1 uncorrelated) for a region centred on Alberta using 00:00 GMT and 12:00 GMT synoptic maps in the period 1946 - 1971. The four most frequent types are shown in Figure 1.42 along with their annual occurrence frequencies. Although Kociuba used these types in a precipitation forecasting scheme, they can also be used to explain regional air pollution levels (Robinson and Boyle 1979). Angle and Gurlay (1983) applied these same types to test for the representativeness of weather during a field test.

1.4.2 Mountain Systems

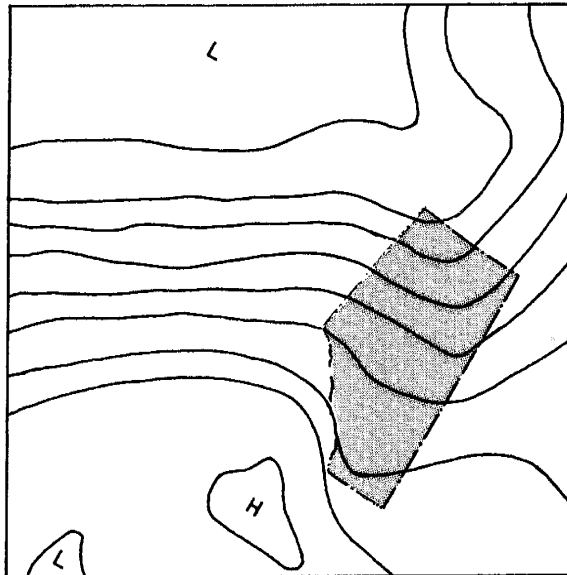
Evidence of mountain winds in Alberta was reported by Longley (1968) who noted that the winds at Calgary showed a marked diurnal variation, being northwest at night and southeast in the afternoon. The same variation extended to altitudes of at least 1000 m, but was absent from surface winds at Coronation, well removed from the mountains. The occurrence of the nocturnal northwesterlies was attributed to an observed diurnal variation in pressure gradient between Banff and Calgary. Munn (1968) explained it as a drainage wind having a trajectory long enough to be affected by the Coriolis force. The resultant flow is then parallel to the Rockies. Longley (1969b) questioned this explanation, noting that sea breezes are affected by the Coriolis force only after 120 miles and that even then the wind does not parallel the coastline. No explanation has been put forward for the afternoon southeasterly wind.



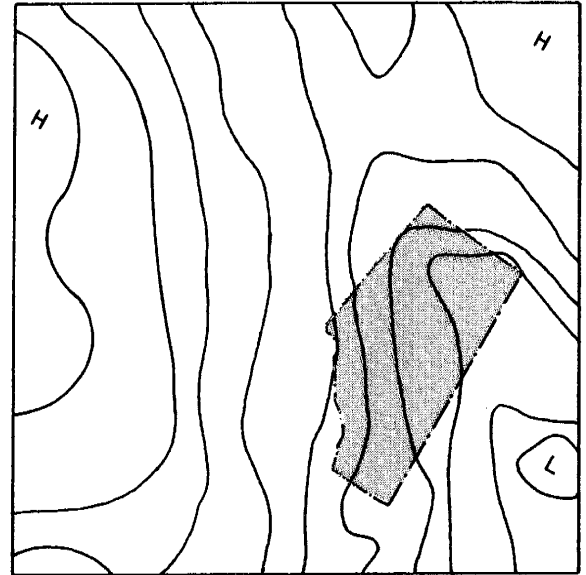
Map type 1 (5.6%)



Map type 2 (7.9%)



Map type 3 (4.5%)



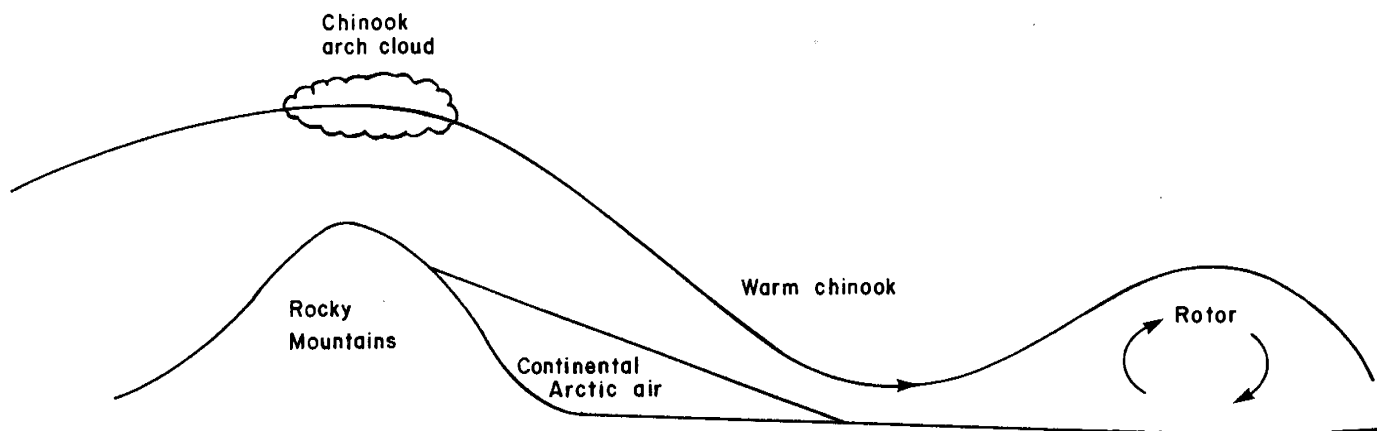
Map type 4 (8.1%)

Figure 1.42 Most frequent Alberta weather map types (Kociuba 1974).

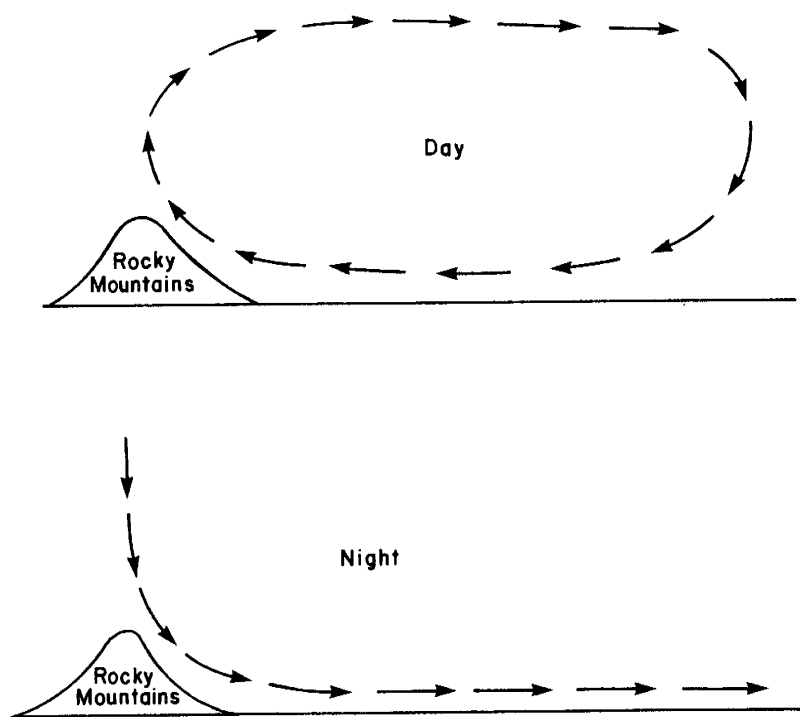
Benjamin (1975) found that the winds at Rocky Mountain House and Red Deer showed broadly the same diurnal variation. Pilot balloon ascents from Calgary revealed that the pattern extends to heights of over 1000 m. Throughout the period of study the large-scale flow was northwesterly as determined by the Stony Plain radiosonde and the upper portions of the pibal ascents. Benjamin speculated that there is a corridor to the east of the Rockies where an organized cellular circulation exists (Figure 1.43). The implication is that plumes would become trapped in a quasi-closed loop and the residence time of pollutants would be greatly increased. Klemm (1977) reported that a weak northwest-southeast pattern is discernible as far east as Wimborne and Nevis. Further evidence of a mountain plain circulation was reported by Thyer (1981) for upper winds and Leahey and Davies (1977) and Western Research (1977) for surface winds.

The most spectacular-mountain related system is the chinook wind. The chinook is commonly defined as a warm, dry, gusty wind from the mountains accompanied by rapid temperature changes. Chinooks occur in a band 100-200 km wide along the eastern slopes of the Rockies from the Yukon to Colorado, but appear to be most intense, frequent and widespread in southern Alberta and northern Montana. Grace (1987) gave an overview of this phenomenon and provided a list of references.

The origin of the warmth of the flow is often explained in geography books as the result of moist adiabatic cooling as the Pacific air rises over the mountains and dry adiabatic warming as the air descends onto the plain. In this classical picture, there must always be precipitation on the windward side of the mountain, a situation not borne out by observations. Since the difference between moist adiabatic and dry adiabatic lapse rates is about $0.004^{\circ}\text{C}/\text{m}$ air rising 3000 m to clear the Rockies and descending to the foothills would experience a warming of about 12°C . This will account for some chinook events, but not those extreme events where temperature increases of up to 25°C in one hour have been observed. An alternative explanation would lie in the dry adiabatic descent only of air from higher altitudes, although some cause for the descent must then be found. Brinkmann (1971) repeated an earlier suggestion that supercritical flow, more commonly called the "hydraulic jump," is one possible mechanism. Cold frontal blocking is another (e.g., Scorer 1974). The presence of a shallow layer of cold Arctic air banked against the mountains is instrumental in accounting for large temperature increases (Glenn 1961).



(a) Chinook with trapped cold pool, long lee wave and shorter wave with rotor.



(b) Mountain-plain circulation

Figure 1.43 Schematic of mountain wind systems.

The effect of a sudden airmass change would then be added to the adiabatic warming effect. At night, shear-generated turbulence prevents strong inversion formation and keeps minimum temperatures relatively high.

Lester (1976a) applied an earlier theory that a chinook is composed to two parts: (1) a macroscale advective component which involves the exchange of airmasses and large-scale subsidence in the lee of the mountains; and (2) a mesoscale wave component which involves gravity waves excited by the mountains and subsequent momentum transfers. The well-known "chinook arch" cloud appears to be a manifestation of long lee waves of wavelength 50 to 100 km. (Typical lee waves have wavelengths of about 10 km). Such wave clouds occur 141 days per year at Pincher Creek, but not all are associated with surface chinooks (Lester 1976d).

Favorable conditions for chinooks are statically stable, southwesterly synoptic flow with a ridge of high pressure on the upwind side of the mountain and a trough of low pressure on the downwind side (the chinook "nose") (Lester 1975; McCabe 1962; Drews 1982; Lester 1976b, c). The onset of the chinook is often marked by low-level speed maxima in wind profiles to the lee of the mountains, reflecting the mesoscale wave component. The transition zone between the warm and cold air weakens as the "chinook front" moves eastward. Chinooks do occur in summer but the temperature contrasts are not as dramatic as in winter (Lester 1976c).

One of the outstanding problems in understanding chinooks lies in finding a satisfactory definition so that results of different workers can be compared and so that the effects can be separated from those of other phenomena (Brinkmann 1971). Ideally the definition would reflect an understanding of the causal mechanisms, but at the present most working definitions are in terms of observed effects. Lester (1976c) defined a chinook event as a period when (1) the daily maximum temperature is above the mean maximum for the month (2) the wind direction is SSW to WNW (3) the wind speed is 4.5 m/s or greater. Five years of Calgary data gave an average frequency of 17.8 (range 2 to 29) chinook events per month for the period October to March and an average of 13.5 days per month (range 2 to 20) with chinook events. Longley (1967) used a winter day with a maximum temperature above 40°F as an indicator of a chinook. The maximum frequency of occurrence was found near Lethbridge, at about 40%. Other preferred areas were in the Calgary-Red Deer corridor and near Entrance. These three "centres of action" lie to the

east of the main gaps in the Rocky Mountains. At the Alberta-Saskatchewan border the frequency of occurrence is less than one-half that in the foothills. Wilson (1973) noted that there is usually a weak trough at the eastern limit of the chinook and that is normally located a little east of Lethbridge, Calgary and Edson.

Hicks and Mathews (1979, 1978) observed that a chinook event produce a recognizable pattern in SODAR records. The unique signature is characterized by (1) a descending multilayered return, accompanied in some cases by wave motion and (2) the formation of a thick (200-500 m) ground-based layer of turbulence having a diffuse and variable upper boundary. The descending structure is the interface between the cold air near the ground and the warm air aloft. The approach of the echo to ground level is well correlated with the onset of westerly winds and surface warming. The turbulent chinook layer can persist for several hours or for several days, long after the strong westerlies have ceased. In the winter of 1976-77 chinook layers were observed 2.7% of the time in November, 17.1% of the time in December and 3.3% of the time in February. Further observations are reported by Mathews et al. (1984) and Leelanda et al. (1981).

The high level of turbulence (Lester and MacPherson 1977) is one of the significant features of chinooks with respect to plume diffusion. Such inordinate rates of vertical spread could lead to a type of fumigation that would produce large concentrations quite close to the stack. The rotor that may develop below the crest of a lee wave (Figure 1.43) is a similar region of high turbulence and, in addition, of reverse flow. The fact that the chinook may not "surface" and that warm air then overrides cold air leads to the formation of an intense chinook inversion (Holmes and Hage 1971). Walker, et al. (1967) observed relatively low values of turbulence intensity under such conditions at Suffield, Alberta. The intensity of vertical turbulence was generally less than that of cross-wind turbulence, and both decreased with height in the lowest 50 to 100 m.

The airflow through a crest towards a trough of a wave has a substantial downward component that could carry a plume to the ground in a large-scale topographic downwash. In the trough itself wind speeds can be very large. Forest "blowdowns" mark regions of preferred positions for wave troughs and evidence their magnitude. Such high winds would tear a plume from a stack and cause impingement quite close to the stack (not to mention the structural stresses on the stack itself). A severe chinook windstorm

that struck southwestern Alberta in October/November 1975 contained hurricane-strength winds over large areas (gusts exceeded 65 m/s) causing extensive timber loss and property damage. Although the wind weakened with distance, gusts exceeded 27 m/s east of Calgary (Lester 1978). Typical chinooks have 8-m/s winds with daily gusts of 17.5 m/s, but about 10% of chinook hours will have winds in excess of 12 m/s and gusts exceeding 23 m/s (Lester 1976c).

1.5 SUMMARY

Wind is the velocity of moving air whose strength and direction are determined by the balance among several forces: pressure gradient, Coriolis, centripetal and friction. Wind can be characterized by concepts such as prevailing direction, persistence, consistency, steadiness or constancy. Wind can be described by statistical distributions such as Weibull or log-normal and can be graphically portrayed in a variety of ways. Persistent light winds are most likely to occur in west-central and northeastern Alberta. Upper level winds tend to be west northwest. Surface winds are dependent upon local terrain and may exhibit a distribution ranging from uniform to a predominant southwesterly flow.

Temperature stratification defines the stability of the atmosphere with respect to displacements of air parcels. Elevated inversions may be caused by fronts, by subsidence, or by differential radiation, mixing, evaporation and condensation in different layers of the atmosphere. Ground-based inversions are the result of radiation cooling or movement over cool surfaces. Inversion strengths, frequencies and depths have been measured near Calgary, Edmonton and Fort McMurray. Ground-based inversions are generally present from sunset to a few hours after sunrise. Gradients vary with the measurement interval, the value $0.08^{\circ}\text{C}/\text{m}$ occurring as a mean in the lowest 20 m and as a 95th percentile in a 100-m layer. Typical inversion depths can vary from a few hundred metres to more than a kilometre. Inversions persist throughout the day perhaps as much as 20% of the time in winter and 5% in other seasons.

The chaotic motions comprising turbulence obtain their energy from either the shear of the mean wind or from solar heating. The nature of turbulence is described by the relative contributions of mechanical and thermal energy. Stability, in this context, refers to the augmentation or suppression of shear-generated turbulence by buoyancy forces.

Turbulence typing schemes try to characterize the thermal component in the absence of heat flux measurements. A variety of such schemes have been applied in Alberta, but it is not clear which one provides the best representation for Alberta's climate and topography.

Synoptic weather patterns influence wind, temperature stratification, and turbulence. The Arctic airmass is associated with a radiation inversion about 1 km deep. The Arctic front is usually present somewhere in Alberta during the winter. Quasistationary fronts create a potential for high ground-level concentrations because light winds occur with a sharp inversion. Mountain wind systems lead to a recirculation zone extending east of the Rockies 150 km and up to 1 km above the ground. Chinooks are associated with elevated inversions and strong winds and turbulence.

1.6 REFERENCES

- Alp, E., M. J. E. Davies, R.G. Huget, L. H. Lam, and M.J. Zelensky. 1990. Gascon 2 A model to estimate ground-level H₂S and SO₂ concentrations and consequences from uncontrolled sour gas releases. prep. for Energy Resources Conservation Board by Concord Environmental Corporation. October 1990. ERCB Report 90-B, 275 pp.
- Anderson, R., B. W. Boville and D. E. McClellan. 1955. An operational frontal contour-analysis model. *Quart. J. Roy. Meteor. Soc.* 81: 588 - 599.
- Angle, R. and P. Gourlay. 1983. Dispersion climatology of the Moose Mountain Area. *Alberta Environment*. February 1983. 58 pp.
- Atwater, M.A. and R.J. Londergan. 1985. Differences caused by stability class on dispersion in tracer experiments. *Atmos. Environ.* 19: 1045 - 1057.
- Barry, R.G. 1967. Seasonal location of the arctic front over North America. *Geographical Bulletin* 9: 79 - 95.
- Benjamin, S.F. 1975. The fate of SO₂ from sour gas plants. UNISUL - Group 4 report. Sept. 1975 Chemical Engineering Dept. University of Calgary. 33 pp.
- Berkowicz, R. and L.P. Prahm. 1982. Sensible heat flux estimated from routine meteorological data by the resistance method. *J. Appl. Meteor.* 21: 1845 - 1864.
- Bowen, B.M., J.M. Dewart, and A. I. Chen. 1983. Stability class determination: a comparison for one site. *Sixth Symp. Turb. Diff.* March 22 - 25, 1983. Boston, Mass. *Amer. Meteor. Soc.* 211-214.
- Briggs, G.A. 1988. Analysis of diffusion field experiments. *Lectures in Air Pollution Modelling*. A. Venkatram and J.C. Wyngaard, ed. *Amer. Meteor. Soc.* 63-117.
- Briggs, G.A. 1982. Simple substitutes for the Obukhov length. *Third Joint Conf. Appl. Air Poll. Meteor.* 12-15. January 1982. San Antonio, Texas. *Amer. Meteor. Soc.* 68-71.
- Briggs, G.A. 1970. Diffusion calculations for small emissions. Contribution No. 36. *Air Resources Turbulence and Diffusion Laboratory*. Oak Ridge, Tennessee, ATDL contribution. 31 pp.
- Brinkmann, W.A.R. 1971. What is a Foehn? *Weather* 26: 230 - 239.
- Brinkmann, W. and I.Y. Ashwell. 1968. The structure and movement of the chinook in Alberta. *Atmosphere* 6:1-10.
- Bryson, R.A. 1966. Air Masses, streamlines and the boreal forest. *Geographical Bulletin* 8: 228 - 269.
- Carpenter, S.B., T.L. Montgomery, J. M. Leavitt, W. C. Colbaugh and F. W. Thomas. 1971. Principal plume dispersion models: TVA power plants. *J. Air Poll. Cont. Assoc.* 21: 491 - 495.

- Choukalos, M. 1985. Summary of data from Tall Meteorological Towers in Calgary and Edmonton. Alberta Environment. March 1985. 89 pp.
- Courtney, F.E. 1977. The determination of temperature lapse rate frequencies in the lower 500 ft. using vertical temperature differences and wind speed relationships. Joint Conf. on Appl. of Air Poll. Meteor. Nov. 29 - Dec. 2, 1977, Salt Lake City, Utah. Amer. Meteor. Soc., Boston. 64-67.
- Davenport, A.G. and C.J. Baynes. 1972. An approach to the mapping of the statistical properties of gradient winds (over Canada). Atmosphere. 10: 80 - 92.
- Davies, M.J.E. and W.C. Thompson. 1983. A comparison of airport and on-site wind and stability data to evaluate sulphur deposition from a sour gas processing plant. PNWIS/APCA, Annual General Meeting, 16 - 18 Nov. 1983, Seattle, Washington.
- DeMarrais, G.A. 1978. Atmospheric stability class determinations on a 481 m tower in Oklahoma. Atmos. Environ. 12: 1957 - 1964.
- Denison, P.J. 1977. A climatology of low-level air trajectories in the Alberta Oil Sands area. Alberta Oil Sands Environmental Research Program. Report 15. Edmonton. 118 pp.
- Draxler, R.R. 1987. Accuracy of various diffusion and stability schemes over Washington, D.C. Atmos. Environ. 21: 491 - 499.
- Draxler, R.R. 1979. Estimating vertical diffusion from routine meteorological tower measurements. Atmos. Environ. 13: 1559 - 1564.
- Drews, M.D. 1982. An aid to forecasting chinooks over southern Alberta. Atmospheric Environment Services Technical Memoranda ISSN0708-2878. Downsview, Ontario. 20 pp.
- EPA (Environmental Protection Agency). 1986. Guideline on Air Quality Models (Revised). Office of Air Quality Planning and Standards, Research Triangle Park, NC 27711. July 1986. 267 pp.
- Fulle, D. 1976. A comparison of three stability classification systems using surface and radiosonde data for four cities in the western United States. Preprint. Third Symp. Atmos. Turb. Diff. and Air Quality. October 19 - 22, 1976. Raleigh, N.C. Amer. Meteor. Soc. 141-148.
- Gifford, F.A. 1976. Turbulent diffusion typing schemes, a review. Nuclear Safety 17: 68 - 85.
- Gildart, M.C. and D.P.Y. Chang. 1983. A temporal comparison of common stability typing methods. J. Air Poll. Cont. Assoc. 33: 491 - 493.
- Glenn, C.L. 1961. The Chinook, Weatherwise 14: 175 - 182.
- Golder D. 1972. Relations among stability parameters in the surface layer. Boundary Layer Meteor. 3: 47 - 58.
- Gorr, W.L. and R.W. Dunlap. 1977. Characterization of steady wind incidents for air quality management. Atmos. Environ. 11: 59 - 64.

- Grace, B. 1987. Chinooks. *Chinook: Summer/87*:52 - 56.
- Hage, K.D. and R.W. Longley. 1970. Ventilation and mixing in Alberta cities. IN: *Man and His Environment*, ed. M. A. Ward. Proceedings of the First Banff Conference on Pollution. 1968 May 16 - 17, Banff, Canada. Pergamon Press. 99-113.
- Hanna, S.R., G.A. Briggs, J. Deardorff, B.A. Egan, F.A. Gifford and F. Pasquill. 1977. AMS workshop on stability classification schemes and sigma curves - summary of recommendations. *Bull. Amer. Meteor. Soc.* 58: 1305 - 1309.
- Hawrelak, J., T. Rachuk and J. Barlishen. 1976. *Wind Power in Alberta*. Alberta Research Council, Edmonton. 110 pp.
- Hayashi, M. 1977. An experimental approach to the Pasquill stability classification and its application to air pollution forecasting. 4th International Clean Air Congress. May 16 - 20, 1977. Tokyo Japanese Union of Air Pollution Prevention Associations. 102-166.
- Heidorn, K.C. 1978. Air pollution incidents and wind variability in southern Ontario. *Atmos. Environ.* 12: 2251 - 2257.
- Hennessey, J.P. Jr. 1977. Some aspects of wind power statistics. *J. Appl. Meteor.* 16: 119 - 128.
- Henry, T.J.G. 1957. Maps of upper winds over Canada. Meteorological Branch, Dept. of Transport. 61 pp.
- Hicks, R.B. and T. Mathews. 1979. Impact of chinooks on Calgary's air quality: Acoustic sounder observations of atmospheric stability. *Water, Air and Soil Poll.* 11: 159 - 172.
- Hino, M. 1968. Maximum ground-level concentration and sampling time. *Atmos. Environ.* 2: 149 - 165.
- Holmes, R.M. and K.D. Hage. 1971. Airborne observations of three chinook-type situations in southern Alberta. *J. Appl. Meteor.* 10: 1138 - 1153.
- Holtstag, A.A.M. 1984. Estimates of diabatic wind speed profiles from near-surface weather observations. *Boundary Layer Meteor.* 29: 225-250.
- Holtstag, A.A.M. and A. P. van Ulden. 1983. A simple scheme for daytime estimates of the surface fluxes from routine weather data. *J. Clim. Appl. Meteor.* 22: 517 - 529.
- Holzworth, G.C. 1974. Climatological data on atmospheric stability in the United States. *Symp. on Atmos. Diff. and Air Poll.*, Sept. 9 - 13, 1974, Santa Barbara, California.
- Holzworth, G.C. 1969. Large-scale weather influences on community air pollution potential in the United States. *J. Air Poll. Cont. Assoc.* 19: 248 - 254.

- Holzworth G.C. 1962. A study of air pollution potential for the western United States. *J. Appl. Meteor.* 1: 366 - 382.
- Irwin, J.S. and F.S. Binkowski. 1981. Estimation of the Monin-Obukhov scaling length using on-site instrumentation. *Atmos. Environ.* 15: 1091 - 1094.
- Islitzer, N. F. and D.H. Slade. 1968. Diffusion and transport experiments. Chapter 4, *Meteorology and Atomic Energy*. D. H. Slade (editor) U.S. AEC.
- James F. MacLaren Limited. 1978. Data source catalogue. Alberta Oil Sands Environmental Research Program. 170 pp.
- Justus, C.G., W.R. Hargraves and A. Yalcin. 1976. Nationwide assessment of potential output from windpowered generators. *J. Appl. Meteor.* 15: 673 - 678.
- Kociuba, P.J., M. Shewel and N. Parker. 1977. Inversion climatology of Edmonton. unpublished.
- Kaimal, J.C., J.C. Wyngaard and D.A. Haugen. 1968. Deriving power spectra from a three-component sonic anemometer. *J. Appl. Meteor.* 7: 827 - 837.
- Klein, W.H. 1957. Principal tracks and mean frequencies of cyclones and anticyclones in the northern hemisphere. Res. Paper No. 40. Weather Bureau. U.S. Dept. of Commerce. 60 pp.
- Klemm, R.F. 1977. Sulphur in Alberta precipitation - consequences of three years of survey. Proceedings of Alberta Sulphur Gas Research Workshop III, Edmonton, Alberta, November 17 - 18, 1977. Alberta Department of the Environment, Research Secretariat, Edmonton. 293-305. See also: Klemm, R. F., 1977. Sulphur in precipitation, contribution to an Alberta sulphur inventory. Alberta Department of the Environment, Edmonton. 126 pp.
- Kociuba, P.J. 1974. Weather map typing and applications for Alberta. M.Sc. Thesis. Meteorology Division, Dept. of Geography, University of Alberta. 217 pp.
- Kretzschmar, J.G. and I. Mertens. 1980. Influence of the turbulence typing schemes upon the yearly average ground-level concentrations calculated by means of a mean wind direction model. *Atmos. Environ.* 14: 947 - 951.
- Lalas, D.P., V. Catsoulis and M. Petrakis. 1979. On the consistency of stability classification schemes when applied to non-homogeneous terrain. *Atmos. Environ.* 13: 687 - 691.
- Leahey, D.M. and M.J.E. Davies. 1977. A preliminary study of the effects of mountain generated wind systems on air pollutant levels in Calgary. Presented at the 1977 Annual Meeting of the Air Poll. Cont. Assoc. - Pacific Northwest Section, November, 1977, Spokane, Washington.
- Leahey, D.M. and M.C. Hansen. 1985. Doppler acoustic radar observations of wind in the Ellerslie region of Alberta. Prep. for Alberta Environment by Western Research, Division of Bow Valley Resources Services Ltd. RMD Report 85/31. 77 pp.

- Leelanda, S.A., S.K. Aggrawal and T. Mathews. 1981. Temperature and velocity turbulence during the onset of chinooks. Proc. Int. Symp. on Remote Sensing of the Atmosphere and Oceans, 22 - 25 June 1981, Calgary, Alberta, VI, 12 - 26.
- Legge, A.H. 1988. The present and potential effects of acidic and acidifying air pollutants on Alberta's environment. Critical Point 1, 1988 Final Report. Prep. for the Acid Deposition Research Program by the Kananaskis Centre for Environmental Research. The University of Calgary, Calgary, Alberta. ADRP-B-16-88. 79 pp. ISBN 0-921625-23-5.
- Lester, P.F. 1976a. Evidence of long lee waves in southern Alberta. Atmosphere 14: 28 - 36.
- Lester, P.F. 1976b. An observational and synoptic study of the chinook arch cloud. Environmental Sciences Centre (Kananaskis). The University of Calgary. 27 pp.
- Lester, P.F. 1976c. A quantitative definition of the chinook in southern Alberta. Environmental Sciences Centre (Kananaskis). The University of Calgary. 28 pp.
- Lester, P.F. 1976d. A lee wave cloud climatology for Pincher Creek, Alberta. Environmental Sciences Centre (Kananaskis). The University of Calgary, Calgary, Alberta. June 1976. 27 pp.
- Lester, P.F. 1975. A case study of the Alberta chinook. Paper presented at the ninth congress of the Canadian Meteorological Society, Vancouver, B.C. May 28 - 30.
- Lester, P.F. and J.I. MacPherson. 1977. Waves and turbulence in the vicinity of a chinook arch cloud. Monthly Weather Review 105: 1447 - 1457.
- Lester, P.F. 1978. A severe chinook windstorm, paper presented at the Conference on Sierra Nevada Meteorology, June 19-21, 1978, South Lake Tahoe, CA. 14 pp.
- Liu, M.K. and D. Durran. 1977. On the prescription of the vertical dispersion coefficient over complex terrain. Joint Conf. on Air Poll. Meteor. Nov. 29 - Dec. 2, 1977. Salt Lake City, Utah. 172-174.
- Longley, R.W. 1969a. The diurnal pressure wave in western Canada. J. Appl. Meteor. 8: 754 - 761.
- Longley, R.W. 1969b. The diurnal variation of wind direction at Calgary: further comments. Atmosphere 7: 35 - 36, 1-2.
- Longley, R.W. 1968. The diurnal variation of wind direction at Calgary. Atmosphere 6: 23 - 29.
- Longley, R.W. 1967. The frequency of winter chinooks in Alberta. Atmosphere 5: 4 - 16.
- Ludwig, F.L. and W.F. Dabberdt. 1976. Comparison of two practical atmospheric stability classification schemes in an urban application. J. Appl. Meteor. 15: 1172 - 1176.

- Luna, R.E. and H.W. Church. 1974. Estimation of long-term concentrations using a "wind universal speed distribution". J. Appl. Meteor. 13: 910 - 916.
- Luna, R.E. and H.W. Church. 1972. A comparison of turbulence intensity and stability ratio measurements to Pasquill stability classes. J. Appl. Meteor. 11: 663 - 669.
- Maarouf, A. and D.B. Fraser. 1972. Estimates of extremes of hourly winds by direction for selected stations in Alberta. Atmospheric Environment Service, Environment Canada. 79 pp.
- Malkiewicz, A.W. and R.V. Portelli, 1976. A Climatological Study of the Air Pollution Potential of the Alberta Oil Sands Region. Atmospheric Environment Service. Report ARQT-3-76, May 1976. 61 pp.
- Mathews, T. and R.B. Hicks. 1984. Doppler sodar measurement of wind climatology of Calgary. Prep. for Alberta Environment, Research Management Division by Department of Physics, The University of Calgary. RMD Report 84/28. 149 pp.
- Mathews, T. and R.B. Hicks. 1979. Typical features of atmospheric turbulence profiles associated with chinooks. Atmospheric Ocean 17: 125 - 134.
- Mathews, T., P.F. Lester and R.B. Hicks. 1984. SODAR observations of chinooks and arctic front passages across the eastern slopes of the Canadian Rockies. Atmosphere Ocean 22: 328 - 342.
- McCabe, H.J. 1962. A new look at chinooks. Dept. Comm. No. 15, Canada Dept. of Transport, Meteor. Branch, Calgary, Alberta. 4 pp.
- MEP Company. 1976. Relations between stability parameters at Mildred Lake. prep. for Syncrude Canada Ltd., November 1976. 35 pp.
- Miller, M.E. and L.E. Niemeyer. 1963. Air pollution potential forecasts - a year's experience. J. Air Poll. Cont. Assoc. 13: 205 - 210.
- Misra, P.K. and S. Onlock. 1982. Modelling continuous fumigation of Nanticoke generating station plume. Atmos. Environ. 16: 479 - 489.
- Mitchell, A.E. Jr. 1982. A comparison of short-term dispersion estimates from various atmospheric stability classification methods. Atmos. Environ. 16: 765 - 773.
- Mitchell, A.E. Jr., and K. O. Timbre. 1979. Atmospheric stability class from horizontal wind fluctuation. Presented at 72nd Annual Meeting of Air Poll. Cont. Assoc., Cincinnati, OH, June 24 - 29, 1979.
- Moses, H. and P.E. Hess. 1976. Comparison of the surface and geostrophic winds. Presented at the 69th Annual Meeting of the Air Pollution Control Association, Portland, Oregon. June 27 - July 1, 1976. paper 76-23.4, 24 pp.

- Munn, T.E. 1968. Note regarding the diurnal variation of wind direction at Calgary. *Atmosphere* 6: 29.
- Munn, R.E., J.Tomlain and R.L. Titus. 1970. A preliminary climatology of ground-based inversions in Canada. *Atmosphere*. 8: 52 - 68.
- Munn, R.E. and J.H. Emslie. 1964. The frequency and intensity of early morning ground inversions at Edmonton 1950 - 1960. *Met. Br., D. O. T., CIR 4009, TEC511*.
- Murray, W. 1983. Statistical analysis of wind and temperature profiles at Mildred Lake. Prep. for Syncrude Canada Ltd. by Promet Environmental Group. December 1983. 33 pp.
- Neumann, J. 1978. Reply. *J. Appl. Meteor.* 17: 1083 - 1084.
- Neumann, J. 1977. Averaging for wind speed in a given direction in the concentration equation for pollutants. *J. Appl. Meteor.* 16: 1097 - 1100.
- Niemeyer, L.E. 1960. Forecasting air pollution potential. *Mon. Wea. Rev.* 88: 88 - 96.
- Ning, D.T. and D. Yap. 1986. Climatology of convective boundary-layer parameters over Ontario, Canada. *Atmos. Environ.* 20: 2315-2323.
- Ontario Ministry of the Environment. 1985. Air quality models suitable for regulatory application. Chapter 12 Air Pollution General Regulation Workshop. November 1985. V-1 to V-40.
- Orlanski, I. 1975. A rational subdivision of scales for atmospheric processes. *Bull. Amer. Meteor. Soc.* 56: 527 - 530.
- Panofsky, H.A. 1969. Air pollution meteorology. *Amer. Scientist.* 57(2):269-285.
- Panofsky, H.A. and J.A. Dutton. 1984. Atmospheric turbulence: models and methods for engineering applications. Wiley Interscience. New York. 397 pp.
- Pasquill, F. 1974. Atmospheric Diffusion. 2nd edition. Halsted Press. John Wiley & Sons, New York. 429 pp.
- Pasquill, F. 1961. The estimation of the dispersion of windborne material. *Met. Mag.*, 90:33-49.
- Pasquill, F. and F.B. Smith. 1971. The physical and meteorological basis for the estimation of the dispersion of windborne material. *Proc. 2nd Int. Clean Air Congress.* Dec. 6-11, 1970. Washington. Academic Press. 1067 - 1072.
- Peake, E., M. A. MacLean and P.F. Lester. 1985. Nitric acid, peroxyacetyl nitrate (PAN), and particulate nitrate in the atmosphere of the City of Edmonton. Prep. for Alberta Environment, Research Management Division by the Kananaskis Centre for Environmental Research, University of Calgary. RMD Report 86/35. 304 pp.
- Pendergast, M.M. and T. V. Crawford. 1974. Actual standard deviations of vertical and horizontal wind direction compared to estimates from other measurements. Preprint Symp. Atmos. Diff. Air Poll. Santa Barbara, Sept. 9 - 13, 1974. *Amer. Meteor. Soc.* 1-6.

- Penner, C.M. 1955. The three-front model for synoptic analyses. *Quart. J. Roy. Meteor. Soc.* 81: 89 - 91.
- Petterssen, S. 1956. *Weather analysis and forecasting*. 2nd ed. Vol. I Motion and Motion Systems. McGraw Hill. 266 pp.
- Plate, E.J. 1971. Aerodynamic characteristics of atmospheric boundary layers. AEC critical review series. US Atomic Energy Commission. 190 pp.
- Portelli, R.V. 1976. A comparative study of experimentally measured atmospheric stability and "STAR program" predictions. *Proc. 3rd Symposium on Atmospheric Turbulence, Diffusion and Air Quality*. Raleigh, N.C. pp. 124 - 129.
- Rao, R.K. and A. Roffman. 1977. Air quality prediction in surface obstructed terrain using different types of stability classification systems. *Joint Conf. Appl. Air Poll. Meteor.* Salt Lake City, UT. Nov. 29 - Dec. 2, 19. Amer. Meteor. Soc. Boston, MA, 221 - 224.
- Reiquam, H. 1980. Stability climatology from on-site wind data. 2nd Joint Conf. *Appl. Air Poll. Meteor.* March 24 - 27, 1980. New Orleans, LA. Amer. Meteor. Soc., Boston. 803 - 808.
- Reynolds, G.W. 1978. Vertical changes in stability conditions within the first 100 meters - Tennessee Valley. Division of Environmental Planning, Tennessee Valley Authority, Muscle Shoals, Alabama.
- Robinson, E. and R.J. Boyle. 1979. Synoptic weather typing and its application to air quality in the St. Louis, Missouri area. 72nd Annual Meeting of Air Poll. Control Assoc., Cincinnati, Ohio June 24 - 29, 1979. paper 79 - 2.4.
- Sakiyama, S. and J. Wong. 1987. Summary of data from the Ellerslie Meteorological Station 1978 - 1985. *Alberta Environment*. October 1987. 120 pp.
- Sandusky, W.F. 1985. Measured atmospheric stability near Oak Ridge, Tennessee. *J. Air Poll. Cont. Assoc.* 35: 121 - 126.
- Schayes, G. 1982. Direct determination of diffusivity profiles from synoptic reports. *Atmos. Environ.* 16: 1407 - 1413.
- Sedefian, L. and E. Bennett. 1980. A comparison of turbulence classification schemes. *Atmos. Environ.* 14: 741 - 750.
- Scorer, R.S. 1978. *Environmental Aerodynamics*. Ellis Horwood Limited. Chichester 487 pp.
- Scorer, R.S. 1974. Seminar on environmental meteorology. University of Calgary, May 23, 1974. 28 pp.
- Shaw, R.W., M.S. Hirt and M.A. Tilley. 1972. Persistence of light surface winds in Canada. *Atmosphere* 10: 33 - 43.
- Shaw, R.W. and R.E. Munn. 1971. *Air pollution meteorology. Introduction to the Scientific Study of Atmospheric Pollution*. B. M. McCormac (ed), D. Reidel Publishing Company, Dordrecht, Holland. 53 - 96.

- Sherlock, R.H. 1951. Analyzing winds for frequency and duration. *Meteorological Monographs*, No. 4, Amer. Meteor. Soc. p. 42 - 49.
- Shirvaikar, V.V. 1972. Persistence of wind direction. *Atmos. Environ.* 6: 889 - 898.
- Sigl, A.B., R.B. Corotis and D.J. Won. 1979. Run duration analysis of surface wind speeds for wind energy application. *J. Appl. Meteor.* 18: 156 - 166.
- Singer, I.A. 1967. Steadiness of the wind. *J. Appl. Meteor.* 6: 1033 - 1038.
- Singer, I.A. and C. M. Nagle, 1970. Variability of wind direction within the United States. *Nuclear Safety* 11: 34 - 39.
- Singer, I.A., J.A. Frizzola, and M.E. Smith. 1966. A simplified method of estimating atmospheric diffusion parameters. *J. Air. Poll. Cont. Assoc.* 16:94-596.
- Singer, I.A. and M.E. Smith. 1953. Relation of gustiness to other meteorological parameters. *J. Meteor.* 10:121-126.
- Skaggs, D.L. and E. Robinson. 1976. A comparison of methods for estimating atmospheric stability and diffusion coefficients. *J. Air Poll. Cont. Assoc.* 26: 888 - 891.
- Slade, D.H. 1968. Meteorological fundamentals for atmospheric transport and diffusion studies. Chap. 2. *Meteorology and Atomic Energy*. U.S. Energy Commission. July 1968. TID-24190. 13-63.
- Smith, F.B. 1979. The relation between Pasquill stability p and Kazanski-Monin stability m (in neutral and unstable conditions). *Atmos. Environ.* 13: 879 - 881.
- Smith, F.B. 1972. A scheme for estimating the vertical dispersion of plumes from a source near ground level. *Proc. Third Meeting Expert Panel on Air Pollution Modelling*, NATO Committee on Challenges of Modern Society. XVII-1 to XVII-14.
- Smith, M.E. 1974. Deficiencies in data and analyses for environmental impact statements. Presented at 67th Annual Meeting of APCA, Denver, Colorado, June 9-13, 1974. paper 74-126.
- Stanford Research Institute, 1958. The uses of meteorological data in large scale air pollution surveys. State of California, Dept. of Public Health, Bureau of Air Sanitation, Berkeley, Calif. June, 1958. 110 pp.
- Steenbergen, J.D. 1971. Comparison of urban and rural turbulence statistics at Edmonton, Alberta. M.Sc. Thesis. Meteorology Division, Dept. of Geography, University of Alberta. 86 pp.
- Stewart, D.A. and O.M. Essenwanger. 1978. Frequency distribution of wind speed near the surface. *J. Appl. Meteor.* 17: 1633 - 1642.
- Sutherland, R.A., F.V. Hansen, and W.D. Bach. 1986. A quantitative method for estimating Pasquill stability class from wind speed and sensible heat flux density. *Boundary Layer Meteor.* 37: 357 - 369.

- Sweeney, J.C. 1979. Inversion persistence at Long Kesh, Northern Ireland. *Weather* 34:50 - 59.
- Tagliazucca, M. and T. Nanni. 1983. An atmospheric diffusion classification scheme based on the Kazanski-Monin stability parameter. *Atmos. Environ.* 17: 2205 - 2211.
- Takle, E.S. and J.M. Brown. 1978. Note on the use of Weibull statistics to characterize wind-speed data. *J. Appl. Meteor.* 17: 556 - 559.
- Thyer, N. 1981. Diurnal variations of upper winds in summer in Alberta. *Atmosphere Ocean* 19: 337 - 344.
- Trombetti, F., M. Tagliazucca, F. Tampieri and T. Tirabassi. 1986. Evaluation of similarity scales in the stratified surface layer using wind speed and temperature gradient. *Atmos. Environ.* 20: 2465 - 2471.
- Turner, D.B. 1964. A diffusion model for an urban area. *J. Appl. Meteor.* 3: 83 - 91.
- Turner, D.B. 1961. Relationships between 24-hour mean air quality measurements and meteorological factors in Nashville, Tennessee. *J. Air Poll. Cont. Assoc.* 11:483-489
- U.S. Nuclear Regulatory Commission. 1972. Onsite meteorological programs. NRC Regulatory Guide 1.23, Washington, D.C.
- van Ulden, A.P. and A.A.M. Holtslag. 1985. Estimation of atmospheric boundary layer parameters for diffusion applications. *J. Clim. Appl. Meteor.* 24: 1196 - 1207.
- Wagonner, P.E. 1965. *Agricultural Meteorology. Meteorological Monographs.* vol. 6, No. 28. Amer. Meteor. Soc., Boston, MA. 188 pp.
- Walker, E.R. 1964. Atmospheric turbulence characteristic measured at Suffield Experimental Station. *Beiträge zur Physik der Atmosphäre.* 37:38 - 52.
- Walker, E.R., O. Johnson and W.H. Boswell. 1968. Meteorological characteristics of chinooks in Alberta. Suffield Technical Note No. 227. Dept. of National Defense.
- Wang, I.T. 1981. The determination of surface-layer stability and eddy fluxes using wind speed and vertical temperature gradient measurements. *J. Appl. Meteor.* 20: 1241 - 1248.
- Wang, I.T. and P.C. Chen. 1980. Estimations of heat and momentum fluxes near the ground. Second Joint Conf. on Appl. of Air Poll. Meteor. March 24 - 27, 1980. New Orleans, LA. Amer. Meteor. Soc., Boston. 764-769.
- Wang, I.T., R. L. Myers and C. M. McKinley. 1978. Concerning the use of routine meteorological data in estimating atmospheric diffusion parameters. *J. Appl. Meteor.* 17:166 - 170.

- Weber, A.H., K.R. McDonald and G.A. Briggs. 1977. Turbulence classification schemes for stable and unstable conditions. Joint Conf. Appl. Air Poll. Meteor., Salt Lake City, Utah, Nov. 29 - Dec. 2, 1977. Amer. Meteor. Soc., Boston, MA. 96 - 102.
- Western Research. 1977. A preliminary study of the effects of the chinook on the air pollutant levels in Calgary. prep. for Alberta Environment March 1977. 59 pp.
- Widger, W.K. Jr. 1977. Estimations of wind speed frequency distributions using only the monthly average and fastest mile data. J. Appl. Meteor. 16: 244 - 247.
- Wilczak, J.M. and M.S. Phillips. 1984. An indirect estimation of convective boundary layer structure for use in routine dispersion models. Fourth joint Conf. on Appl. Air Poll. Meteor. 16 - 19 Oct. 1984, Amer. Meteor. Soc., Boston. 16-19.
- Williamson, H.J. and R.R. Krenmayer. 1980. Analysis of the relationship between Turner's stability classifications and wind speed and direct measurements of net radiation. Second Joint Conf. on Applications of Air Pollution Meteorology. March 24-27, 1980. New Orleans, La. Amer. Meteor. Soc., Boston. 777-780.
- Wilson, E. 1979. Regional Frequency Distributions of Mixing Depth Parameters in Canada. A Climatological Study of Air Pollution Potential. Atmospheric Environment Service Report ARQN-3-79, July 1979.
- Wilson, H.P. 1978. Chinook airflow. In Essays on Meteorology and Climatology, K.D. Hage and E.R. Reinelt (Eds), University of Alberta Press, Edmonton, Alberta. 387 - 397.
- Wilson, H.P. 1973. Inertia lee waves and weather in Alberta. Presented at CMOS Alberta Centre meeting. 11 pp.
- Wratt, D.S. 1987. An experimental investigation of some methods of estimating turbulence parameters for use in dispersion models. Atmos. Environ. 21: 2599 - 2608.

2.0 THE ATMOSPHERIC BOUNDARY LAYER

The *atmospheric boundary layer* (ABL) or *planetary boundary layer* (PBL) is defined as that portion of the atmosphere which is directly influenced by the earth's surface through the vertical exchange of momentum, heat and moisture. With a strong wind, the thickness of the PBL is the height at which the Ekman spiral gives way to the geostrophic wind. When there is an upward heat flux from the surface, the boundary layer thickness is the height at which the potential temperature gradient returns to the stable stratification of the free atmosphere; in other words the thickness is determined by the height of an elevated capping inversion. If the atmosphere is unstable over a deep layer so that convective clouds and rain showers develop, the thickness of the PBL is poorly defined and seems to include the entire troposphere. If cloud development is limited and no precipitation occurs, cloud base may be taken as the PBL thickness. When the heat flux is downward, turbulence is suppressed, the coupling between layers is weakened, and it becomes difficult to define the PBL. The situation is often complicated by the presence of gravity waves, intermittent "bursts" of turbulence and changes in wind shear. In Arctic winters, situations arise where the winds are light and the temperature does not vary appreciably from the surface to 10 km.

The PBL actually consists of two layers, each governed by a different set of flow parameters. The surface layer, which is typically 20 to 200 m thick, has its structure determined by the constant fluxes of heat and momentum which in turn depend on the nature of the ground. The outer layer is a transition to the free atmosphere.

2.1 THE NEUTRAL BOUNDARY LAYER

The structure of the neutral boundary layer is depicted schematically in Figure 2.1. Near-neutral boundary layers occur with strong winds and cloudy skies. In neutral conditions all turbulence is generated by wind shear. In the surface boundary layer:

$$\frac{du}{dz} = \frac{u_*}{kz}$$

where u_* is the friction velocity, z is height above ground and k is von Karman's constant. The wind profile is obtained by integration which yields,

$$u = \frac{u_*}{k} \ln \frac{z}{z_0}$$

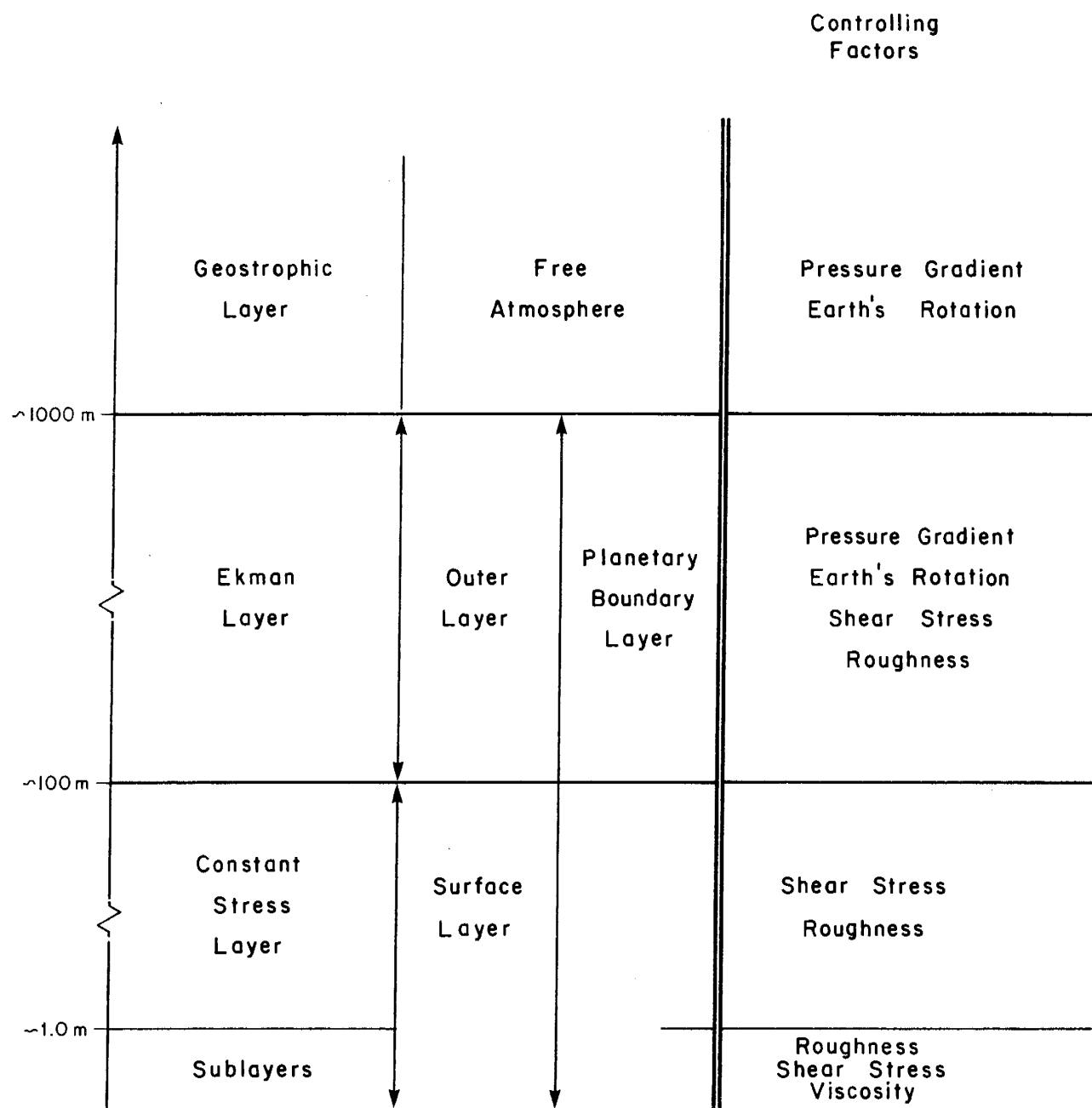


Figure 2.1 Schematic of the neutral atmospheric boundary layer.

where z_0 enters as the constant of integration (the value of z where $u = 0$) but can be identified as the roughness length characterizing underlying terrain. This is often called the *law of the wall* by analogy with wind tunnel flows over flat plates. This logarithmic wind profile applies only down to a height of some multiple of the roughness length below which the air is directly influenced by local pressure forces set up by the individual roughness elements. It has been verified experimentally at numerous instrumented towers around the world and has been found to hold up to 150 m even in situations where the shear stress decreases significantly (Panofsky 1974).

The roughness length and the friction velocity are thus the two parameters governing the wind profile. Measurements of the wind speed at two heights in neutral conditions allow the determination of both. The wind speed required is actually the longitudinal component u , not the horizontal speed u_h measured by a cup anemometer. In practice the two are often used interchangeably although considerable error can result (Bernstein 1967). The error can be corrected using the standard deviation of wind direction fluctuations.

A less accurate alternative relies on the wind speed at one height and an estimate of the roughness length based on the nature of the underlying terrain. A summary of experimentally determined values from around the world is given in Table 2.1. Using pilot balloon measurements of wind up to 150 m Rudolph and Davison (1985) found a roughness of 1 m in the Athabasca River valley north of Fort McMurray and a roughness of 10 m on the neighboring plain. Sakiyama and Wong (1987) observed a roughness of 0.04 m and an average u_* of 0.35 m/s southwest of Edmonton. Many workers have found that the roughness length is about one-fifth the height of the roughness elements. Consideration of the spacing and shape of the roughness elements leads to (Lettau 1969)

$$Z_0 = 0.5 h_e (A_s / A_e)$$

where h_e is the height of the roughness element, A_s is the silhouette area of the average element (cross-sectional area "seen" by the wind in an approach to the obstacle), and A_e is the specific area of the element (average area occupied by a single roughness element). Verified by bushel basket experiments its accuracy was considered to be $\pm 25\%$ and its validity restricted to situations where $A_s < A_e$. Angle (1975) reported good agreement between this "roughness-element description" formula and the observed roughness of a mustard field in southern Alberta.

Table 2.1 Summary of experimentally determined roughness lengths.

Type of Surface	z_o (m)
Forests	1.0 - 10.0
Orchards	0.5 - 2.0
Cities	1.0 - 2.0
Corn	0.7 - 1.3
Wheat	0.03 - 0.23
Alfalfa	0.025 - 0.027
Tall grass	0.039 - 0.15
Short grass	0.014 - 0.032
Lawn	0.001 - 0.005
Root crops ⁰	0.10 - 0.14
Beets	0.064
Mustard	0.10 - 0.14
Fallow	0.01 - 0.02
Snow	0.001 - 0.005
Mountains	3.0 - 3.5

The scatter of values may, in part, be due to the dependence of ground texture upon wind speed in the case of rather pliant crops, since the roughness depends on fairly fine details of surface structure (Monin and Yaglom 1971). Natural vegetation, even of one type, is unlikely to have identical structure in different parts of the world.

When the roughness elements are large an additional parameter is often introduced so that the equation becomes

$$u = \frac{u_*}{k} \ln \frac{z-d}{z_0} \quad \text{for } z > d + z_0$$

where d is the zero-plane displacement. There has been considerable controversy over the exact role of this parameter. Angle (1975) summarized various interpretations of it as: (1) an allowance for the virtually stationary layer of air trapped within the roughness elements, (2) a datum level above which normal turbulent exchange takes place, (3) a shift in the origin of the z -axis because a vegetation cover produces an effective surface above the solid ground, (4) a corrective height increment from which an observer has measured anemometer heights, (5) a translation of the zero-level for the vertical axis analogous to the displacement thickness in fluid mechanical boundary layer theory. The zero-plane displacement is often taken as equal to the height of the roughness elements, but roughness element spacing is obviously important and values as small as one-fifth the height have been reported. Roughness-length and zero-plane displacement are closely related and subject to similar variabilities from similar sources.

For the temperature

$$\frac{d\theta}{dz} = \frac{T_*}{kz}$$

where θ is the potential temperature, and T_* is a scaling temperature given by

$$T_* = \frac{-H}{\rho c_p u_*} = \frac{-(u'T')_{avg}}{u_*}$$

In the outer layer, the turning of the wind with height due to the Coriolis force and the pressure gradient as represented by the geostrophic wind must both be incorporated. It turns out that the appropriate velocity scale is still u_* but the appropriate length scale is u_*/f which is proportional to the height of the boundary layer. At the top of the boundary layer, the geostrophic wind G (with components u_g and v_g) will make an angle α with the surface shear stress. The *surface Rossby number*, G/fz_0 , expresses the relative magnitude of the inertia forces as compared to the Coriolis forces. The ratio of shear stress velocity to geostrophic wind, u_*/G , is called the *geostrophic drag coefficient*.

The logarithmic law is not merely a property of the neutral surface layer, but is also a property of the lower part of the Ekman layer. A logarithmic profile may be expected up to about one-tenth the height of the boundary layer, irrespective of the decrease of stress with height or the turning of the stress (wind direction) with height (Tennekes 1973).

Turbulence and other characteristics of the neutral boundary layer are summarized in Table 2.2.

2.2 THE UNSTABLE OR CONVECTIVE BOUNDARY LAYER

In the surface layer in non-neutral conditions

$$\frac{du}{dz} = \frac{u_*}{kz} \phi_m$$

$$\frac{d\theta}{dz} = \frac{T_*}{kz} \phi_h$$

where L is the Monin-Obukhov length, ϕ_m is the non-dimensional wind shear and ϕ_h is the non-dimensional potential temperature gradient. For unstable conditions $z/L < 0$

$$\phi_m = (1 - \gamma z/L)^{-1/4}$$

a functional form determined entirely from experimental data with $\gamma \approx 15$. Integration then yields a profile of the form

Table 2.2 Summary of equations describing the neutral atmospheric boundary layer
 $(-0.2 \leq h/L \leq 0.2)$.

Property	Surface Layer	Outer Layer
Wind	$u = \frac{u_*}{k} \ln \frac{z}{z_0}$	$u = u_g + (u_*/k)(z/h + A)$ $A \approx 0$ (Panofsky and Dutton 1984)
Temperature stratification	$\theta = \theta_r + (T^*/k) \ln z/z_r$ $T^* = \frac{-H}{\rho c_p u_*}$	$\theta \approx \text{constant}$
Turbulence	$\sigma_w = 1.25 u_*$ $\sigma_v = 1.9 u_*$ $\sigma_u = 2.4 u_*$ (Panofsky and Dutton 1984) $\lambda_{mw} = 5.9z$ $\lambda_{mv} = 5z$ $\lambda_{mu} = 15z$ (Pasquill and Smith 1983)	$\frac{\sigma_w}{u_*} = 1.3 \exp(-2 fz/u_*)$ $\frac{\sigma_{u,v}}{u_*} = 2.0 \exp(-3 fz/u_*)$ $\lambda_m = \frac{5z}{1 + 15 fz/u_*}$ (Hanna et al. 1982)
Layer height	$h_{sfc} = 0.1 h_{pbl}$	$h_{pbl} = 0.17 u_*/f$ (Panofsky and Dutton 1984)

$$u = \frac{u_*}{k} \left(\ln \frac{z}{z_0} - \psi \right)$$

where ψ is the diabatic influence function. Other information is given in Table 2.3.

In mid-latitudes over land the outer layer reaches a height of 1000 - 2000 m by mid-afternoon. Its upper limit is often marked by a sharp "capping" inversion which separates the turbulent boundary layer from the free atmosphere above. Convection produces strong vertical mixing which leads to a near-constant distribution of wind and potential temperature. The structure of the convective boundary layer is shown schematically in Figure 2.2.

In the free convection layer, which starts when $z > -L$, all properties depend upon only two parameters, heat flux and height. Shear stress is not a controlling variable. In the mixed layer above 10% of the depth of the PBL, the properties all scale with heat flux and boundary layer height (Holtslag and Nieuwstadt 1986). With low solar radiation and high winds a near-neutral upper layer will be present ($-h/L < 5$). At the top of the boundary layer, there is a zone called the entrainment layer where air from the free atmosphere mixes with the air in the boundary layer. This layer can be 20 to 700 m thick (Boers and Eloranta 1986). The different regions, together with the appropriate scaling quantities are shown in Figure 2.3.

In the mixed layer the velocity scale is given by

$$w_* = \left(\frac{gHh}{c_p \rho T} \right)^{1/3}$$

where H is the heat flux and h is the boundary layer height. The temperature scale is given by

$$T_* = (w' T')_{\text{avg}} / w_* \quad \text{where } (w' T')_{\text{avg}} = \frac{H}{\rho c_p}$$

Practical estimates for the convective velocity scale w_* can be obtained from (Venkatram 1978):

$$\begin{aligned} w_* &= AH^{1/2} \\ &= B h \end{aligned}$$

Table 2.3 Summary of equations describing the unstable atmospheric boundary layer.

Property	Surface Layer	Convective (Mixed) Layer
Wind	$u = (u_*/k)(\ln z/z_0 - \psi_m)$ $\psi_m = 2 \ln [0.5 (1 + 1/\phi_m)]$ $+ \ln [0.5 (1 + 1/\phi^2 m)]$ $- 2 \tan^{-1} (1/\phi m) + \pi/2$ $\phi_m = (1 - 16 z/L)^{-1/4}$ (Panofsky and Dutton 1984)	$u = \text{constant}$ $\partial u / \partial z = 0$
Temperature stratification	$\theta = \theta_r + (T^*/k)(\ln z/z_r - \psi_h)$ $\psi_h = 2 \ln [0.5 (1 + 1/\phi_h)]$ $\phi_h = (1 - 16 z/L)^{-1/2} = \phi^2_m$ (Panofsky and Dutton 1984)	$\theta = \text{constant}$ $\partial T / \partial z = \gamma_d = 0.01^\circ\text{C/m}$
Turbulence	$\sigma_w = 1.25 u_* (1 - 3 z/L)^{1/3}$ $\sigma_v = u_* (12 - 0.5 h/L)^{1/3}$ (Panofsky and Dutton 1984) $\lambda_{mw} = \frac{z}{(.55 + .38 z/L)}$ $\lambda_{mv} = \lambda_{mw} = 1.5 h$	$\sigma_w = 0.745 W_*/(1 + 0.255(h/z)^{0.5})^{2/3}$ $\sigma_{u,v} \approx 0.58 W_*$ (Pasquill and Smith 1983) $\lambda_{mw} = 1.5 h [1 - \exp(-5z/h)]$ $\lambda_{mu} = \lambda_{mv} = 1.5 h$ (Hanna et al. 1981)
Layer height	$h_{sfc} = 0.1 h$	$h^2 = \frac{2 \int H dt}{\rho c_p \partial \theta / \partial z}$ $H = \text{heat flux}$ $t = \text{time}$ (Panofsky and Dutton 1984)

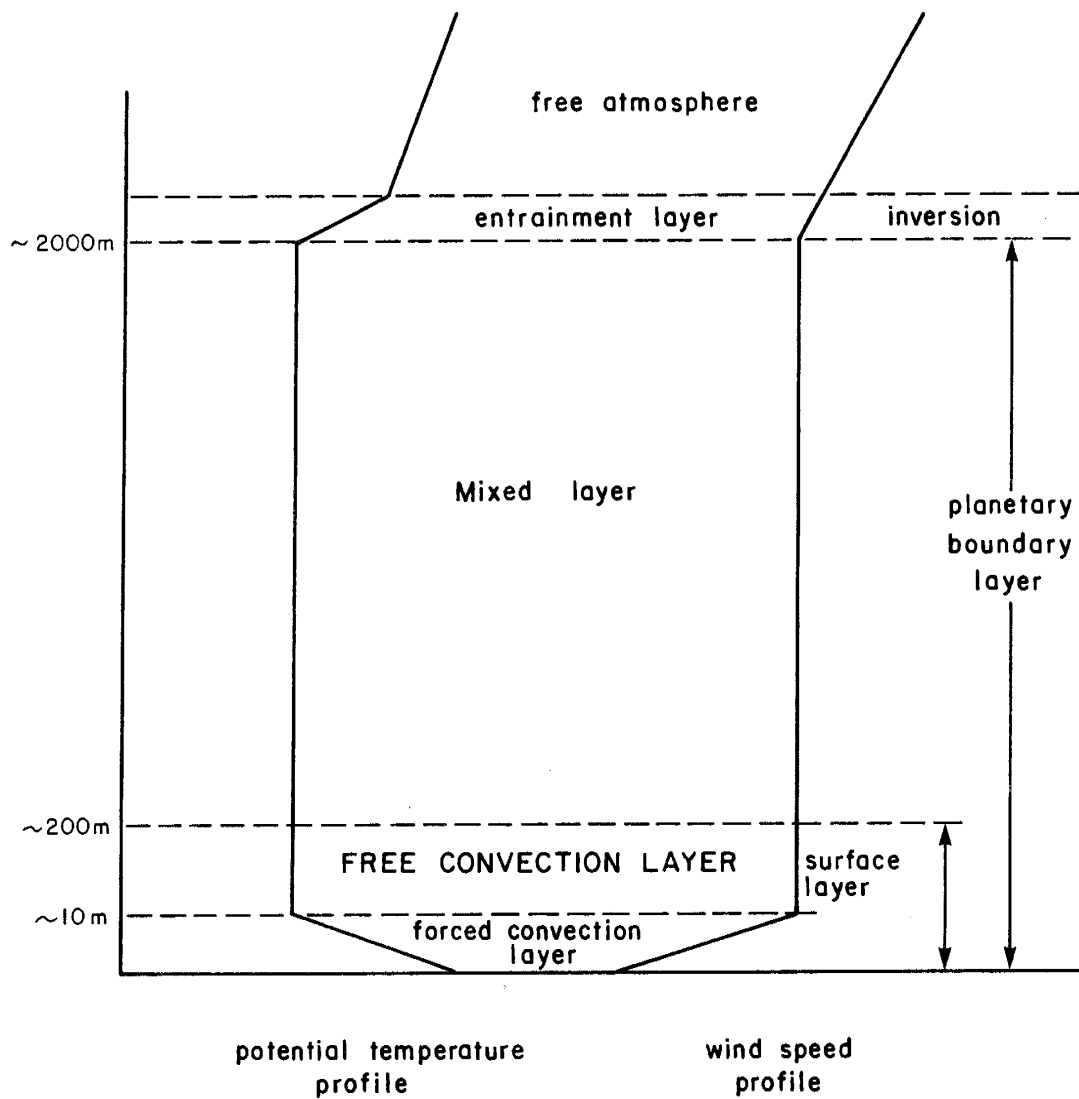


Figure 2.2 Schematic of the unstable (convective) atmospheric boundary layer.

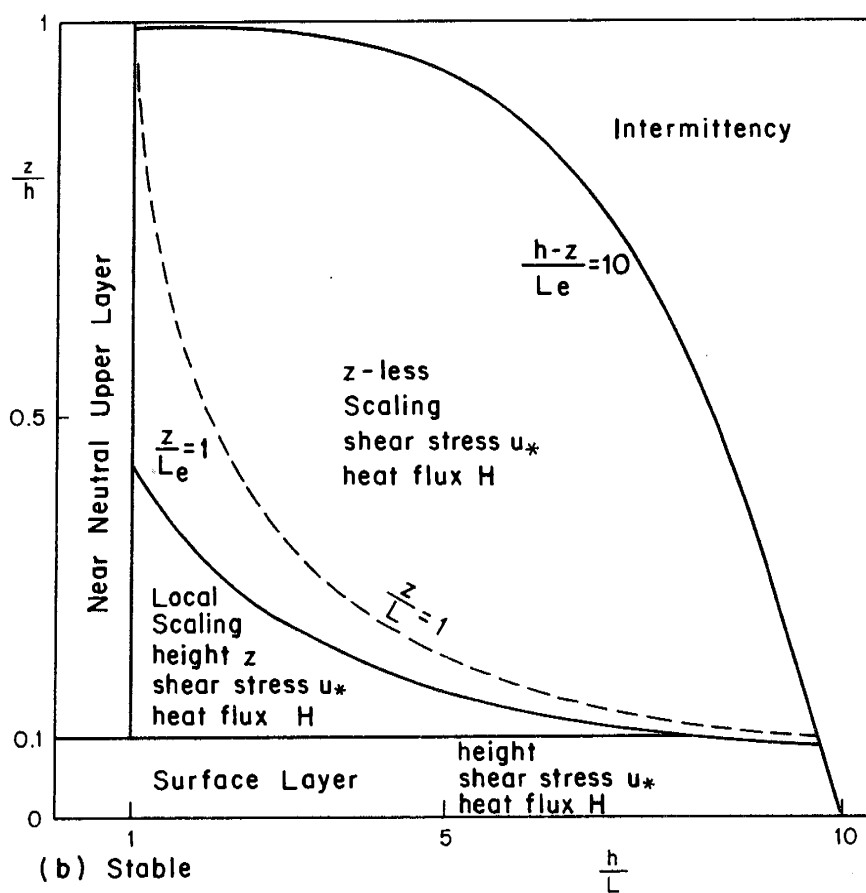
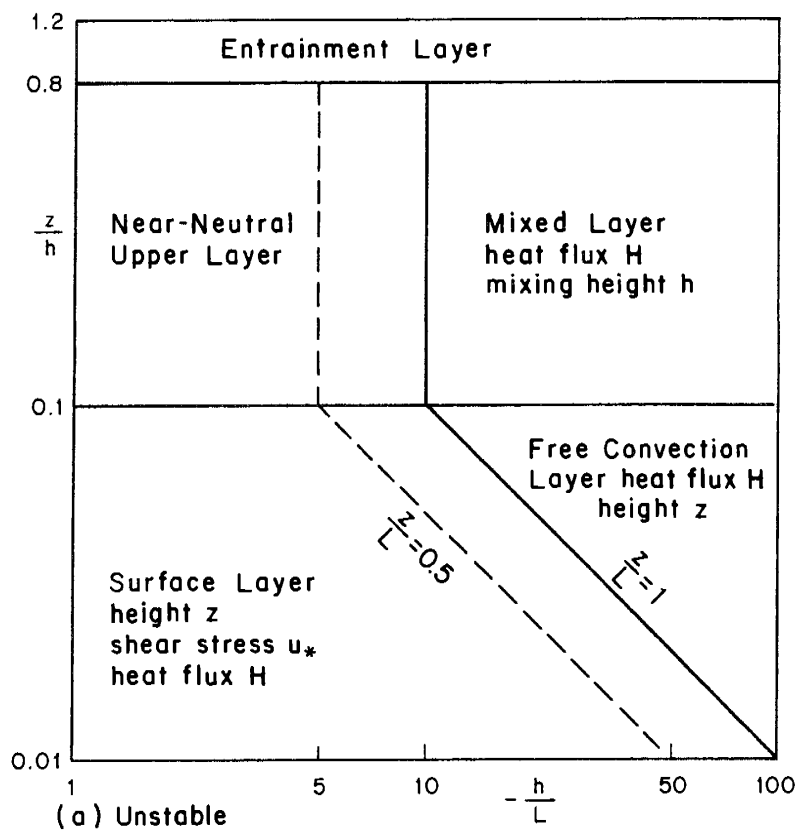


Figure 2.3 Scaling regions for the atmospheric boundary layer (after Holtslag and Nieuwstadt 1986).

Venkatram (1980) assigned the values $A = 4.8$, $B = 1.07 \times 10^{-3}$ for northeastern Alberta in the months May through September. Using estimated heat fluxes he calculated w_* values of 1.5 m/s (7/10 cloudiness) to 2.4 m/s with corresponding mixed-layer heights of 1100 m and 1800 m.

Turbulence characteristics in the unstable boundary layer are summarized in Table 2.3. In the mixed layer there is a strong tendency for the peak wavelength in all components to be the same. The convective field resembles "thermal streets" with individual thermals lined up in the direction of the mean wind spaced roughly 1.5 h apart with a row separation of about 3 h (Kaimal et al. 1976). With lighter winds the pattern can change to open rings with diameter of 5 - 6 h.

The complete spectra in the unstable mixed layer indicate that the inertial subrange exists for wavelengths $\lambda < 0.1$ h and that the energy in the inertial subrange remains constant with height. This contrasts with the sharp decreases observed near the ground. Thus the dissipation rate is also constant with height. Evidence also indicates the presence of a spectral gap between three-dimensional boundary layer turbulence and quasi-horizontal synoptic flows (Kaimal et al. 1976). Energy from disturbances such as thunderstorms and frontal passages would fill the gap, but the convective boundary layer, by most definitions, would not exist during such periods.

The capping inversion results from turbulent mixing between the air in the convection layer and the warmer air in the stable free atmosphere above. This interfacial layer is highly convoluted as it undergoes deformation by thermals carrying heated air upwards. The thermals or convective eddies arrive at the top of the mixed layer with a positive vertical velocity as a result of acceleration during the upward journey. Hence they penetrate into the stable free atmosphere until increasingly negative buoyancy brings them to a halt. Some of the kinetic energy of the thermal is converted to potential energy in that the warmer air is displaced downward in fulfillment of continuity requirements; however, a substantive part becomes turbulent kinetic energy generated as the thermal flattens and the now cooler air falls through the warmer air to its equilibrium position. An inversion is usually regarded as a quiescent layer devoid of turbulence; however, at the top of the convective boundary layer turbulent mixing actually creates the inversion and the

layer is by no means quiescent. Constant agitation by the thermals produces strong mixing which entrains air from the stable region into the top of the convective layer, a mechanism which allows the top of the convective layer to move upward. This agitation also means that the structure of the inversion layer will vary substantially in space and time. Thus the layer cannot always be clearly defined from observations.

This interfacial mixing produces a downward heat flux in the upper portion of the mixed layer. Typically the turbulent heat flux (H) in the mixed layer decreases linearly with height, and reaches zero at about $0.6 h$ (Caughey and Kaimal 1977), thereafter becoming negative until a maximum negative value is reached at the base of the capping inversion. From the base of the inversion to the top of the inversion, the heat flux increases to the value zero which exists in the stable free atmosphere. The value of the ratio between the downward heat flow at the base of the capping inversion and the upward heat flux from the surface is generally taken to be 0.2 ± 0.1 (Caughey and Palmer 1979, Tennekes 1973, Yamada and Berman 1979, Driedonks 1982) although values ranging from 0 to 1 have been reported (Dubosclard 1980). This implies that the entrainment process is relatively inefficient and that many thermals do not break into the stable region, but apparently "bounce off" the inversion. Typical values for the downward heat flux are in the range $30 - 170 \text{ W/m}^2$ (Caughey and Palmer 1979).

2.3 THE STABLE BOUNDARY LAYER

A stable boundary layer is formed during the night when the surface is cooled by long-wave radiation to space. Its structure is shown schematically in Figure 2.4. In the surface layer, for stable conditions

$$\begin{aligned}\psi_m &= -\beta z/L \\ \phi_m &= (1 - \psi_m)\end{aligned}$$

where $\beta \approx 5$

This relationship is taken to mean that in stable air the dominant eddy size becomes independent of height and is proportional to the Monin-Obukhov length.

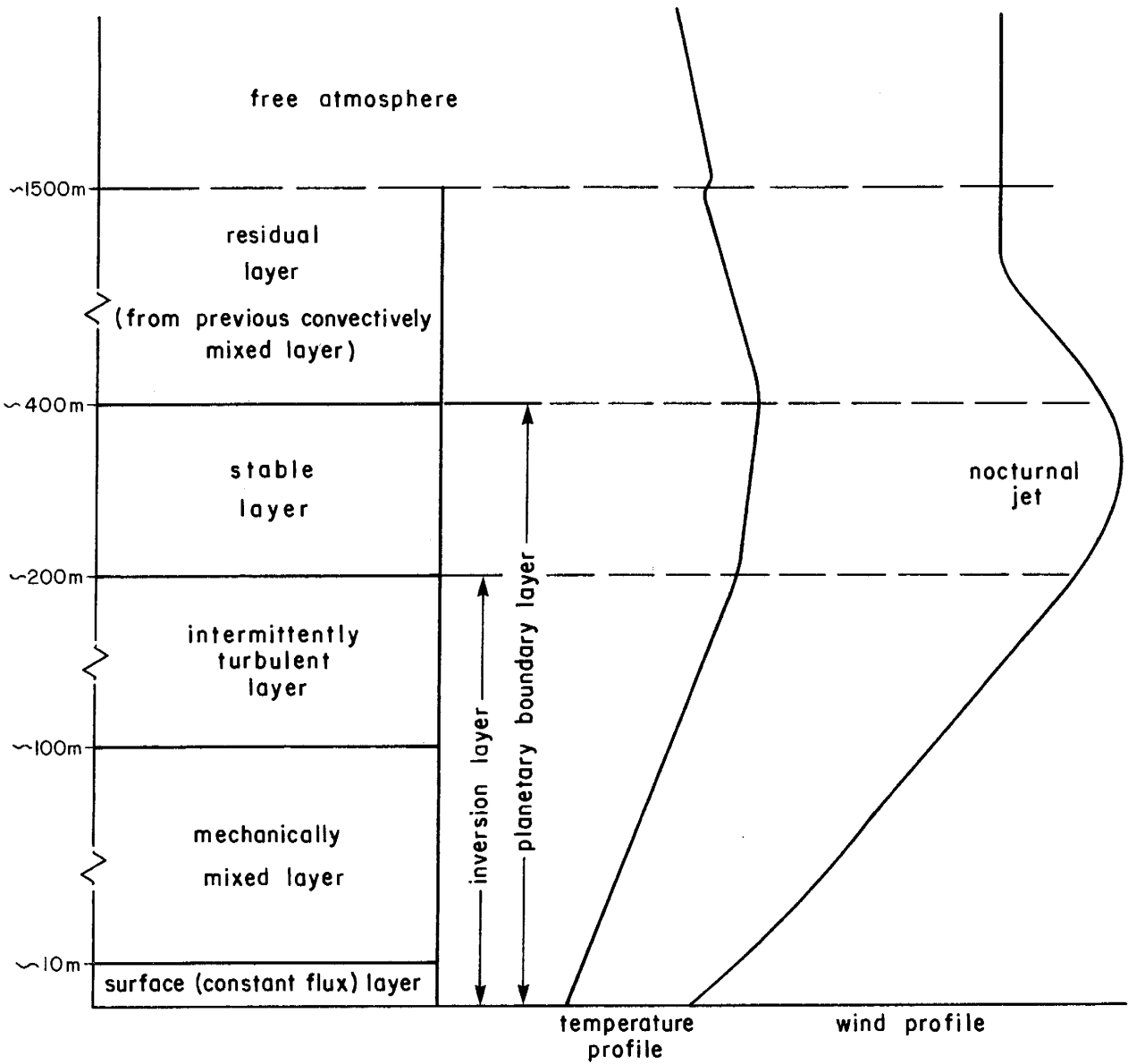


Figure 2.4 Schematic of the stable atmospheric boundary layer.

The foregoing equations can be valid for only a limited range of z/L because they imply a unique relationship between z/L and Ri . The existence of a critical Ri requires either that z/L also take a critical value or that the similarity arguments break down. Generally the limits for this regime may be taken as $0 < z/L < 1$. In strong stability ($z/L \approx 1$) it has been argued that the gradients should be independent of height.

There is also a suggestion that variables not included in the Monin-Obukhov similarity theory may become important (Panofsky 1974); for example, radiative flux may become competitive with turbulent heat flux. In general, wind profile theory is unlikely to be satisfactory above the surface layer. The turning of the wind begins early and as the critical Richardson number is approached and exceeded, winds at different heights are virtually independent of each other. Each level is free to seek a geostrophic equilibrium. Theories based on turbulent mixing simply do not apply in such a situation. However, it has been found (van Ulden and Holtslag 1985) that in strong stability

$$\psi_m = -17 [1 - \exp(-0.29 z/L)]$$

describes the observation quite well. Other characteristics of the stable boundary layer are summarized in Table 2.4.

Under strongly stable conditions turbulence decays and can become intermittent even near the surface (Chiba and Kobayashi 1984, Kondo et al. 1978, Oke 1970). At an Ontario site, it was found that turbulence begins to decay at $Ri = 0.1$ and is very small at $Ri = 0.3$ (Oke 1970); however, a complete cessation of turbulence does not occur and the vertical intensity of turbulence remains at about 1%. In the Antarctic a minimum vertical intensity of 1.5% for values of Ri up to 12 and roughness of 0.24 cm. at a height of 30 m above the ground has been reported (Chiba and Kobayashi 1984). In Japan (Kondo et al. 1978) the vertical intensity was found to be constant for $Ri < 0.7$ and then drop suddenly to one-half the value for $Ri > 2$. The fraction of time with turbulence (defined at standard deviations above a threshold) becomes very small for $Ri > 2$.

During stable conditions at Porton, England, σ_v is insensitive to wind speed and stability, maintaining a constant value of about 0.3 m/s (Hanna 1983). At a rugged site in California (Hanna 1981) a constant value of $\sigma_v = 1$ m/s was found. These values were attributed to a mesoscale turbulence component (meandering) on a time scale of about 2 h. Leahey et al. (1988) reported that σ_v took a constant value of 0.55 m/s on level prairie near Ellerslie, Alberta; 0.56 and 0.65 m/s for prairie sites in the lee of the

Table 2.4 Summary of equations describing the stable atmospheric boundary layer.

Property	Surface Layer	Outer (Mixed) Layer
Wind	$u = u_*/k (\ln z/z_o - \psi_m)$ $\psi_m = -5z/L$ (Panofsky and Dutton 1984)	$u = u_*/k (\ln z/z_o - \psi_m)$ $\psi_m = -17 [1 - \exp (-.29 z/L)]$ (van Uden and Holtslag 1985)
Temperature stratification	$\theta = (T_*/k)(\ln z/z_o - \psi_h)$ $\psi_h = -5 z/L = \psi_m$ (Panofsky and Dutton 1984)	$\theta = T_*/k (\ln z/z_o - \psi_h)$ $\psi_h = -17 [1 - \exp (-.29 z/L)]$ (van Uden and Holtslag 1985)
Turbulence	$\sigma_w = 1.3u_*$ $\sigma_u = 2.4 u_*$ $\sigma_v = 1.9 u_*$ (Panofsky and Dutton 1984) $\lambda_{mw} = \frac{z}{(0.5 + z/L)}$ $\lambda_{mu} = 15 z$ $\lambda_{mv} = 5 z$ (Pasquill and Smith 1983)	$\sigma_w = 1.3 u_* (1 - z/h)^{1/2}$ $\sigma_{u,v} = 2.0 u_* (1 - z/h)^{1/2}$ (Hanna et al 1982) $\lambda_{mu} = 1.5h (z/h)^{1/2}$ $\lambda_{mv} = 0.7h (z/h)^{1/2}$ $\lambda_{mw} = h(z/h)^{0.8}$ (Hanna et al 1982)
Layer height	$h_{sfc} = 0.1 h$	$h = 0.4 (u_* L/f)^{1/2}$ (Panofsky and Dutton 1984)

Rockies and 0.73 m/s for a foothills site. The vertical intensity was found to be 6% at the prairie site and 14% at the foothills site. Alp et al. (1990) used a minimum value of 0.058 m/s for u^* at Calgary which corresponds to a minimum vertical turbulence intensity of 7.5%.

Above the surface layer, a local Obukhov length, L_ℓ , defined in terms of the local heat flux (which decreases with height) will describe the structure (Nieuwstadt 1984a). Above $z = L_\ell$, the dependence on height z vanishes and all dimensionless quantities approach a constant value, for example, $Ri = 0.2$. Above $z = 6L$ turbulence becomes intermittent, that is, weak and sporadic. No longer continuous in space or time, the turbulence is confined to isolated patches which develop and disappear. The scaling regions are shown in Figure 2.3.

The spectrum of vertical velocity has a maximum at shorter wavelengths than in neutral conditions. Thus λ_m/z is smaller than in neutral conditions. There is no obvious effect of stability on the u -component. The v -component appears to have energy damped in the region where neutral spectra have a maximum. This leaves the spectrum containing more energy at both larger and smaller wavelengths. The high-frequency energy decreases rapidly with height, but the low-frequency energy is nearly constant (Pasquill 1974).

The wind speed profile at night is frequently observed to contain a sharp maximum usually located at the top of the nocturnal inversion. This low-level jet contains wind speeds 30 - 60% higher than the geostrophic wind speed appropriate to the synoptic pressure gradient and is often associated with large values of wind shear. This phenomenon has been explained (Blackadar 1957) as an inertial oscillation of the geostrophic defect when the frictional constraint imposed by daytime mixing is released by the initiation of a ground-based radiation inversion (Figure 2.5). The nocturnal jet promotes a gradual and orderly upward growth of turbulence by wind shear.

A number of features of the nocturnal jet have been elucidated (Thorpe and Guymer 1977):

- (1) It occurs over rough terrain as well as smooth.
- (2) Skies must be clear for half an inertial period (a few hours) to allow radiational cooling to set up a stable layer near the ground.

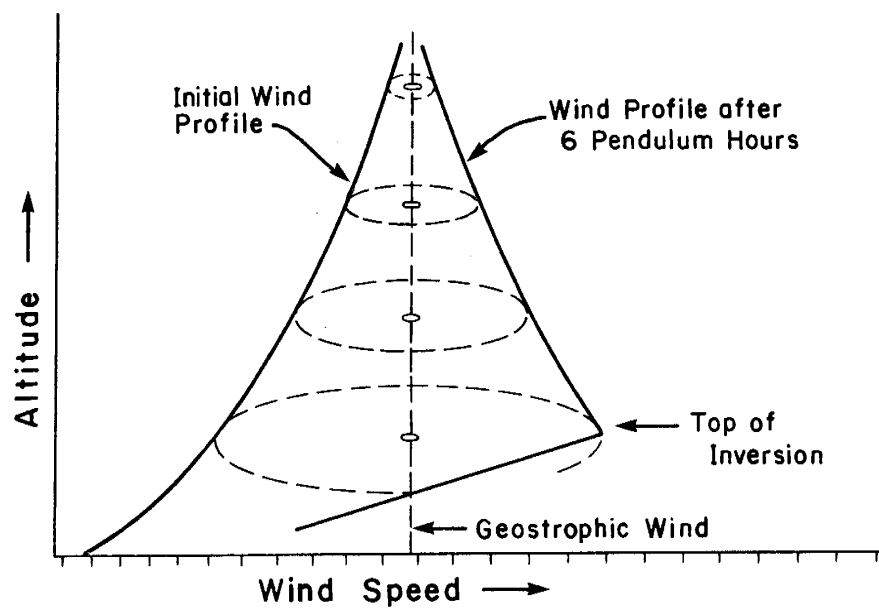


Figure 2.5 Schematic illustration explaining the evolution of a boundary layer jet profile (Blackadar 1957).

- (3) The magnitude of geostrophic defect during the afternoon of the previous day is likely to determine the magnitude of the jet. The jet will not be pronounced if the wind is light or so strong that static stability cannot inhibit the vertical mixing.
- (4) If the geostrophic wind decreases with height (cold air to the right) the jet will be enhanced.
- (5) When the length of night is close to half an inertial period the maximum jet speed will be reached near dawn. At latitude 50° half an inertial period is about 7.5 h which means that a dawn maximum can be expected during spring and autumn.
- (6) The conditions for the appearance of the nocturnal jet are most likely to be satisfied in an anticyclonic region with a substantial pressure gradient.

Such a jet could cause an abrupt termination of the plume rise from an industrial stack. Transport would take place much more quickly than surface winds would indicate.

2.4 THE THICKNESS OF THE PLANETARY BOUNDARY LAYER

The height of the top of the boundary layer is often obtained from temperature soundings and defined as the lowest level at which the vertical temperature gradient exhibits a discontinuity. Under convective conditions this is the base of an elevated inversion or stable layer. In stable conditions this is the top of the ground-based inversion. Discontinuities in wind speed and humidity usually occur near this height as well. However, the thickness of the turbulent region may be quite different, about 10% more in the unstable PBL and considerably less in the stable PBL (Panofsky and Dutton 1984). The observed variation of the boundary layer thickness is well described by the formula of Laikhtman (Hanna 1969):

$$h = CG / \left(\frac{g}{T} \frac{\Delta\theta}{\Delta z} \right)^{1/2}$$

where G is the geostrophic wind, C is a constant equal to about 0.75, and $\Delta\theta/\Delta z$ is the average potential temperature gradient in the boundary layer determined as

$$\frac{\Delta\theta}{\Delta z} = (\theta_h - \theta_{100}) / (h - 100)$$

where θ_{100} refers to the potential temperature at a height of 100 m.

This method can be used most of the time because on the average the atmosphere is statically stable throughout the boundary layer. It implies that the bulk Richardson number equals 1.7 for all stable boundary layers; in other words, the boundary layer somehow adjusts itself to this value.

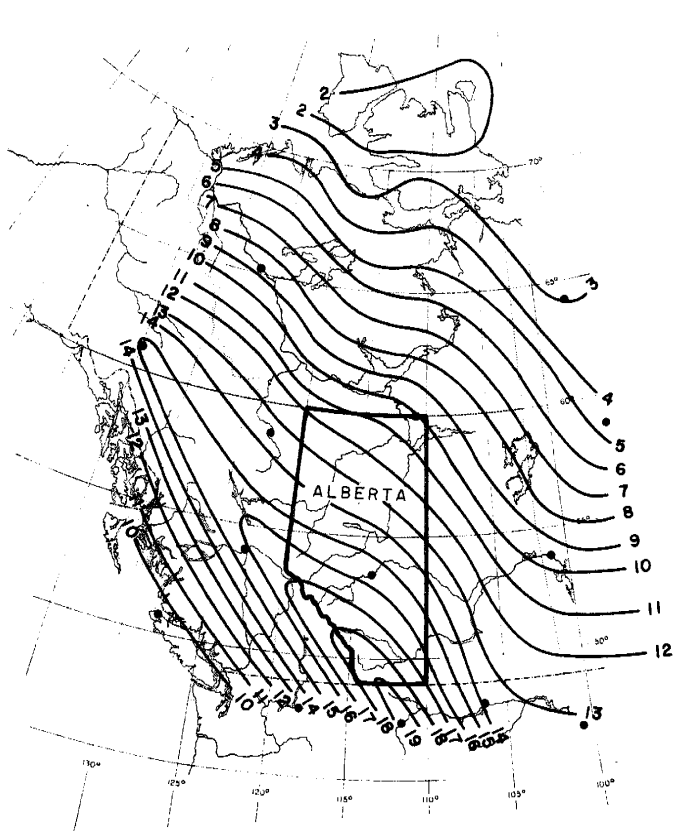
The daily maximum mixing height may be estimated from the morning temperature profile (12 Z radio sounding) using a thermodynamic chart and drawing a dry adiabat upward from the maximum surface temperature for the day (Holzworth 1967). The point of intersection between the adiabat and the morning temperature profile defines the daily maximum mixing height. This method is based on the principle that surface heat flux results in vigorous vertical mixing that leads to the dry adiabatic lapse rate. Although the lapse rate near the ground is usually superadiabatic in sunny conditions, near the time of the maximum surface temperature a quasi-equilibrium is reached and the adiabatic approximation is reasonable (Holzworth 1974). Good agreement is obtained between observed heights and those estimated by the Holzworth method (Hanna 1969).

The distribution of mean seasonal daily maximum mixing heights for Edmonton as given by Portelli (1977a) are shown in Table 2.5. The pattern of mixing heights throughout the province is given in Figure 2.6 from Portelli (1977b).

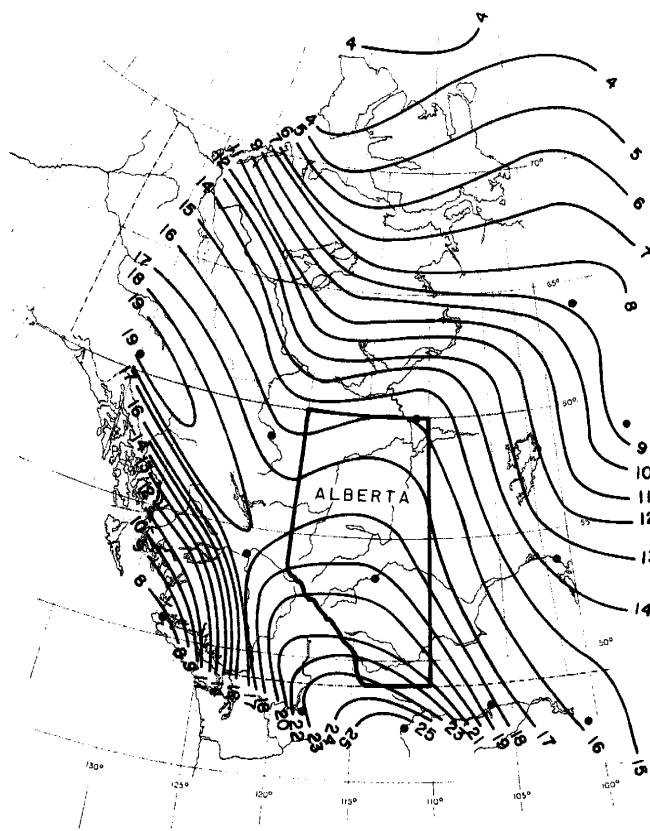
Later workers adapted the procedure for the estimation of hourly mixing heights using hourly surface temperature. Due to the presence of the superadiabatic layer near the ground, one might expect that serious overestimation would occur. However, this is offset to some extent by the deepening attributable to penetrative convection and the maintenance of a capping inversion. This scheme can be improved by incorporating the effects of temperature advection (airmass warming or cooling) and allowing either mechanical or convective daytime mixing (Benkley and Schulman 1979).

Table 2.5 Mean maximum (afternoon) mixing heights at Edmonton (Stony Plain) and the frequency of occurrence in specified intervals, March 1966 to June 1969 (Portelli 1977a).

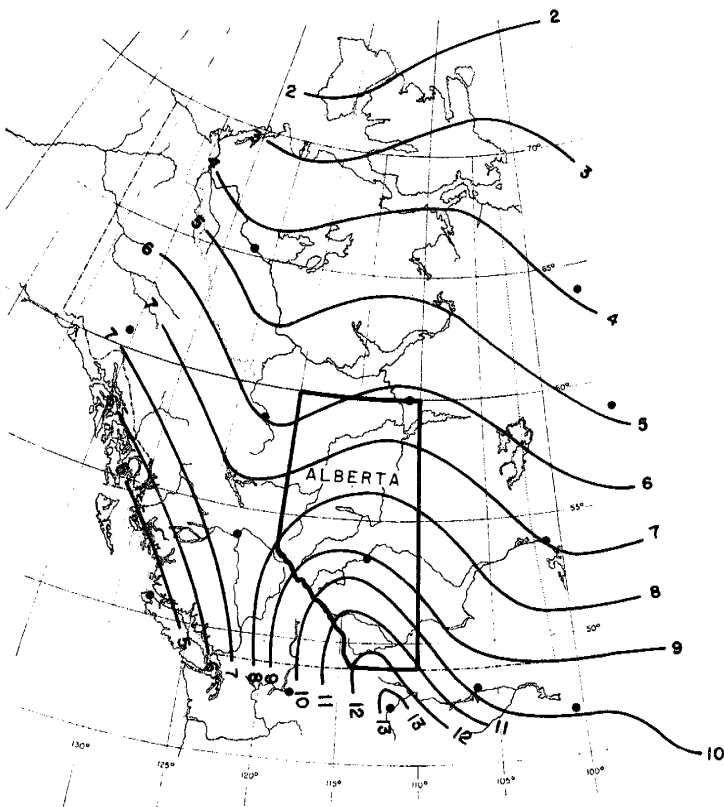
SEASON		Winter	Spring	Summer	Autumn
MEAN	(m)	242	1557	1901	913
	No mixing	14.4	3.0	1.0	3.7
	001 to 199	43.5	6.0	1.8	14.4
	200 to 399	27.3	8.9	3.2	16.5
	400 to 599	10.7	9.8	4.5	11.3
	600 to 799	2.6	8.4	4.4	9.5
	800 to 900	0.8	5.7	3.5	8.8
	1000 to 1199	0.4	7.1	6.6	5.2
	1200 to 1399	0.4	3.3	6.5	5.8
	1400 to 1599	0.0	3.5	5.0	5.8
	1600 to 1799	0.0	4.3	6.1	4.4
	1800 to 1999	0.0	4.1	7.2	4.8
	2000+	0.0	35.9	50.3	9.9



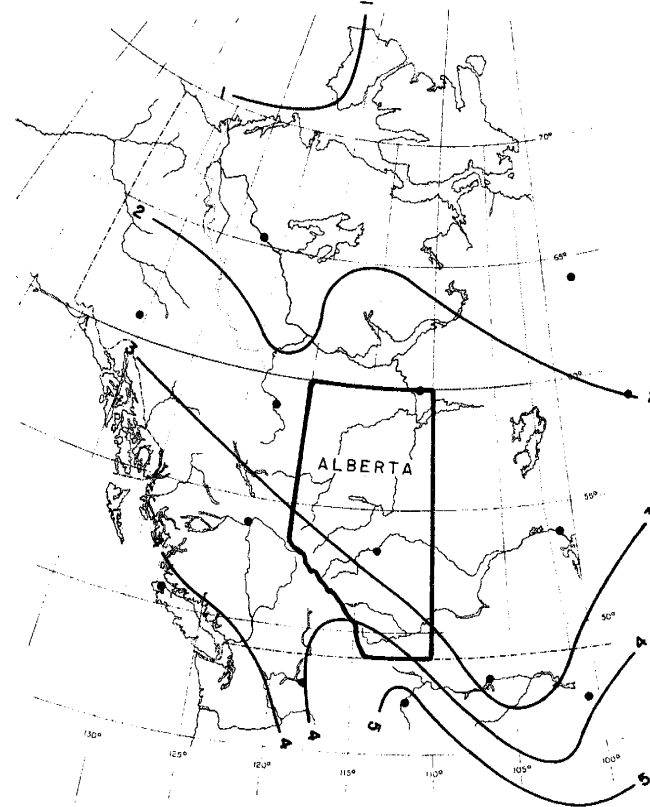
Spring



Summer



Autumn



Winter

Figure 2.6 Isopleths of mean maximum afternoon mixing heights in hundreds of metres (Portelli 1977b)

In another adaptation of the Holzworth technique, the mixing height for a given temperature profile is determined by drawing the dry adiabat upward from the actual surface temperature to its intersection with the rest of the profile. Such a technique (sometimes called the "conventional" method) is extremely sensitive to errors in surface temperature, especially in daytime soundings when the profile tends to be near adiabatic over a substantial height interval. Kumar (1979, 1978c) compared boundary layer heights determined by this method with those defined by the "kink" method, where the boundary layer thickness is taken to be the lowest maximum in the rate of change of temperature with height (effectively the base of the capping inversion). He found that the two were poorly correlated with a linear correlation coefficient of only 0.22 in winter. In general, mixing heights for northeastern Alberta were underestimated in winter and overestimated in other seasons. Aircraft observations in Florida (Sholtes 1972) showed that this technique generally underestimated mixing heights determined by turbulence or temperature. Davison et al. (1981) summarized numerous soundings in the Fort McMurray area using the conventional method (Figure 2.7). Mixing heights near Edmonton using the kink method are summarized in Table 2.6.

Thomson (1977) found that the Holzworth method underestimated mixing heights in a set of minisoundings from the Edmonton and Fort McMurray areas. He also developed an energy balance model to forecast hourly mixing height. The model did not require an initial temperature profile but used as inputs: latitude, longitude, elevation, surface and cloud albedos, minimum and maximum temperatures, temperature and height of the 85- and 70-kPa levels.

In stable conditions defining the boundary layer becomes problematical (Yu 1978). Ideally, the height of the stable boundary layer would be the height at which turbulence disappears and above which shear stress and heat flux become negligible. In practice the stable boundary layer height is defined as: (1) The height of the lowest definite maximum in the wind profile, (2) the height of the nocturnal temperature inversion, (3) the height to which significant cooling has extended (as judged both from individual potential temperature profiles and their evolution in time). Field data from Wangara, Australia revealed that the nocturnal boundary layers for heat and momentum are comparable, but quite different from the height defined by the top of the inversion. All three practical definitions overestimate the actual stable boundary layer height. A summary of various possible definitions of boundary layer thickness is given in Table 2.7.

Table 2.6 Median values of kink mixing height at Edmonton (Ellerslie).

Time of Day	WINTER		SPRING		SUMMER		FALL	
	Median	No. of Obs.	Median	No. of Obs.	Median	No. of Obs.	Median	No. of Obs.
0600					72	32		
0730			216	58	128	93	153	12
0900	282	31	315	82	414	86	108	49
1030	167	56	706	17	513	22	216	13
1200	162	49	659	77	683	80	381	90
1330	216	29	706	13	686	17	398	8
1500	222	41	769	55			579	74

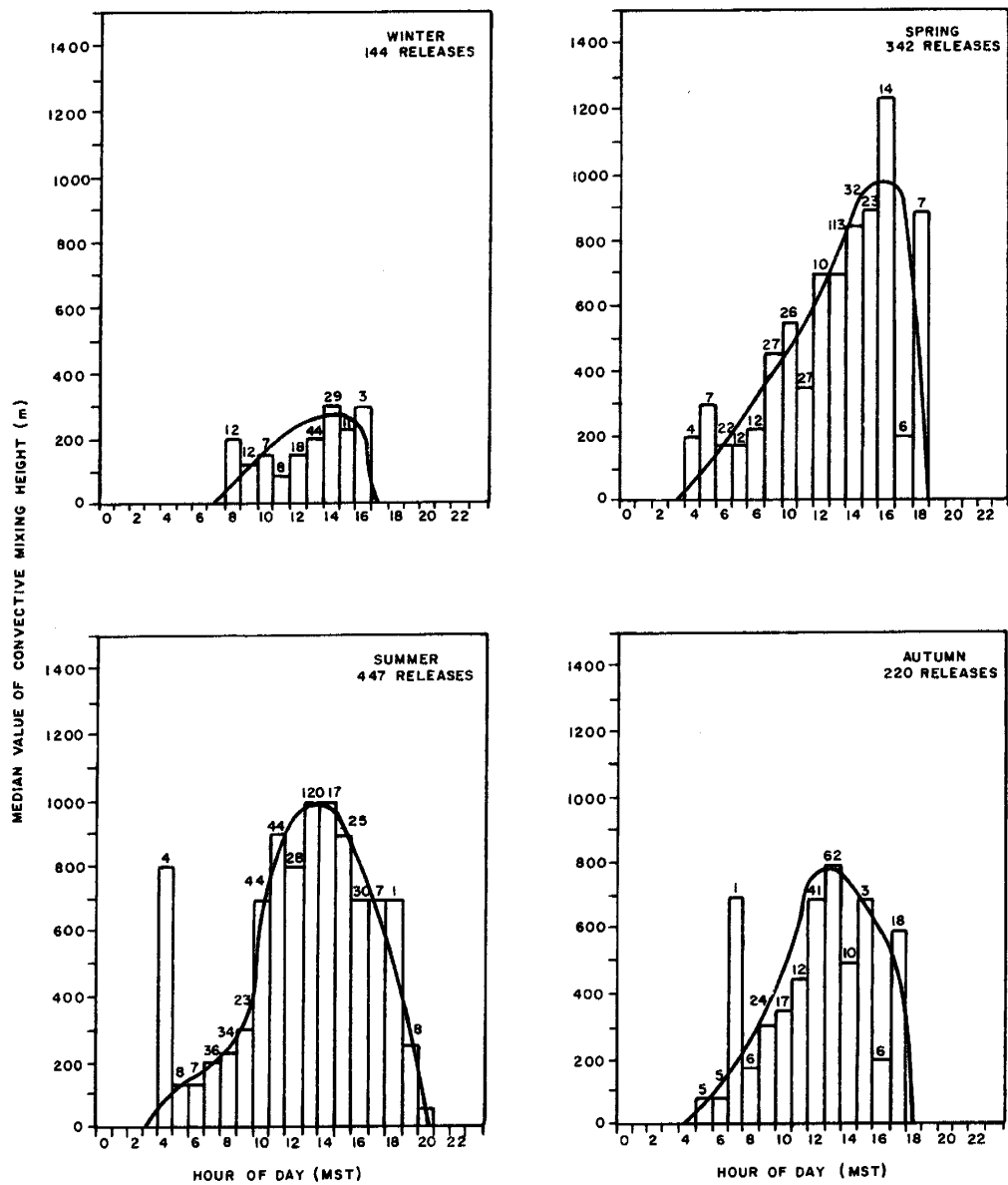


Figure 2.7 Hourly median convective mixing heights at Fort McMurray (Davison et al. 1981).

The measurement methods summarized in Table A.5 introduce several operational definitions.

The top of the layer of significant turbulent fluxes is often found to occur just below the low-level wind maximum, which coincides with a maximum gradient Richardson number (Mahrt et al. 1979), but is 25% lower than the top of the inversion. This has been attributed to clear-air radiational cooling and/or shrinking of the layer of turbulence with time. Operational estimates of the depth of mechanical mixing in a nocturnal inversion have been obtained from $h = 125 u_{10}$ in the U.S. assuming 5-cm roughness where u_{10} is the wind speed at 10 m (Benkley and Shulman 1979) or from $H = 28 u_{10}^{3/2}$ in Holland (Nieuwstadt 1984b) or $0.14 u_*/f$ (Arya 1981). Values measured by acoustic sounder near Edmonton averaged 96 m in July and 127 m in January (Sakiyama and Wong 1987).

Jarvis (1968b) developed a regression equation for predicting the stable boundary layer height (ground-based inversion) using data for a 3-month winter period at Suffield, Alberta:

$$h = 120 + 30 \Delta u + 5 G$$

where h is the boundary layer height in feet, G is the geostrophic wind in mph and Δu is the difference in wind speeds in mph between the geostrophic wind and the wind at 10 m. Testing on independent data yielded a linear correlation coefficient of 0.78 and a root mean square error of 347 ft. (106 m).

Seale (1964) analyzed wind profiles obtained by two-theodolite tracking of pilot balloons near Camp Shilo, Manitoba. Vector averages over five stations on five autumn days gave the depth of frictional influence in morning as about 600 m and in the afternoon as about 900 m.

2.5 POWER LAW FORMS OF THE WIND PROFILE

The dependence of the wind on height is often more conveniently represented by the empirical expression

$$u = u_T (z/z_T)^P$$

Table 2.7 Alternative definitions of the thickness of the atmospheric boundary layer (after Hanna 1969 and Yu 1978).

Thickness Criterion	Definition of h	Interpretation
displacement	$\int_0^{\infty} (1 - \frac{u}{G}) dz$	$G \cdot h$ is the decrease in volume flow due to friction
momentum	$\int_0^{\infty} (\frac{u}{G} (1 - \frac{u}{G})) dz$	$G^2 \cdot h$ is the loss of momentum in the boundary layer
cross-isobar displacement	$\int_0^{\infty} v/G dz$	$G \cdot h$ is the volume flow across the isobars
cross-isobar momentum	$\int_0^{\infty} v/G^2 dz$	$G^2 \cdot h$ is the momentum flow across the isobars
wind direction	$D(h) = DG$	first height where wind assumes the direction of the geostrophic wind
temperature	$\frac{\partial^2 T}{\partial z^2} (h) = 0$	first extreme in the vertical temperature profile (top of ground-based inversion or base of an elevated inversion)
wind speed	$\frac{\partial^2 u}{\partial z^2} (h) = 0$	nocturnal jet height
turbulence	$H(h) = 0$ $\tau(h) = 0$	negligible heat and momentum flux (< 1% surface value)

where u_r is the wind at a reference height z_r and p is a function of stability, roughness and height interval. This empirical equation is useful because it yields a reasonable average representation of the wind profile over the entire boundary layer and because integral relations based upon this simple form are not far from correct (Plate 1971). A nomogram relating p to the roughness length and reciprocal of Monin-Obukhov length for the layer between 11 and 46 m (Panofsky et al. 1960) and tabulations of the power law exponent as a function of roughness and Pasquill stability category for the layer 10 m to 100 m (Irwin 1979) are available, as are early empirical results (DeMarrais 1959).

A summary of p values measured in Alberta and in similar terrain elsewhere in Canada is given in Table 2.8. Pettitt and Root (1965), Wilson (1959) and Jarvis (1968) reported that the values decrease with increasing wind speed. Wilson (1959) found that the angle between the surface wind and the gradient wind averaged 20° at Edmonton and 27° at Fort Smith, while Jarvis (1968) reported a Canada-wide mean of 45° .

2.6 BOUNDARY LAYER EVOLUTION

On a clear night the earth's surface radiates heat to space and cools the layer of air next to it. The transfer of heat from the air to the ground takes place by molecular conduction through a very thin layer because there can be no air velocity at the boundary. Above this thin layer the heat is transferred by mechanical turbulence. As the stable layer thickens its upper part becomes dominated by the downward heat flux until the critical flux Richardson number is exceeded and the turbulence is damped out by the negative buoyancy effect. This effectively "uncouples" the stable surface layer from the flow above it. The stable layer is quickly retarded by friction at the ground which is transmitted throughout the layer by the shear stresses. The layer directly above undergoes an acceleration because it is no longer subject to surface friction. After a short time the velocity gradient everywhere in the lower layer will be reduced so much that the flux Richardson number exceeds its critical value everywhere. Thus turbulence is suppressed and the flow tends towards laminar. The retardation of air in the lower layer and the acceleration of air in the upper layer leads to a decrease in the gradient Richardson number across the interface and allows turbulence to become re-established. Too much turbulence would reduce the gradient and force the flux Richardson number above its critical value.

Table 2.8 Summary of the power law exponents observed in and near Alberta.

Reference	Stability/Time	Terrain/Location	Layer (m)	Value of P
Davenport (1960)	neutral	prairie grassland	PBL	0.14
	neutral	parkland	PBL	0.29
	neutral	large cities	PBL	0.40
Pettitt and Root (1965)	inversion	Whiteshell, Manitoba	6-60	0.40 to 0.24
	lapse	Whiteshell, Manitoba	6-60	0.27 to 0.16
Buckler (1969)	summer	Regina, Saskatchewan	0.5-27	0.25 ± 0.27
	winter	Saskatoon, Saskatchewan	10-27	0.21
Djurfors (1971)	October	Calgary Bonnybrook	10-100	0.45
	July/Jan.	Calgary Bonnybrook	10-100	0.31
Johnson (1959)	neutral	prairie at Suffield	0.5-16	0.22
	inversion	prairie at Suffield	0.5-16	0.49
	lapse	snow at Suffield	0.5-16	0.16
	inversion	snow at Suffield	0.5-16	0.28
Wilson (1959)	annual	Edmonton	PBL	0.21
	annual	Fort Smith	PBL	0.26
Jarvis (1968)	annual	across Canada	PBL	0.14-0.40
Leahey and Hanson (1985)	annual	SW of Edmonton	10-500	0.17 ± 0.12
	annual	SW of Edmonton	150-500	0.30 ± 0.39
	unstable	SW of Edmonton	10-500	median 0.10
	neutral	SW of Edmonton	10-500	median 0.17
	stable	SW of Edmonton	10-500	median 0.28
Murray and Morrow (1977)	stable	north of Fort McMurray	100-2000	0.25
	unstable		100-2000	0.18
Davison et al. (1981)	annual	north of Fort McMurray	400 m - 1200 m	-0.35 to 0.50
Young and Wilson (1980)	annual	across Alberta	10 - PBL/2	0.12 to 0.13

Hence the interface is likely to contain weak turbulence, enough to allow the stable layer to grow, but not enough to eliminate the velocity gradient.

When the lower layer tends to be laminar, the effect of friction is greatly reduced. The Coriolis force leads to a slow acceleration of the flow, taking several hours to reach geostrophic (Plate 1971). As this takes place the gradient Richardson number may again fall below its critical value, especially if there is a change in heat flux, and the lower layer will become turbulent again. Thus the nocturnal boundary layer is subject to irregular "bursts" of momentum and heat. These intermittent episodes of turbulent mixing may occur several times throughout the night.

The depth of the inversion layer generally increases with time (Mahrt et al. 1979, Yamada 1979, Nieuwstadt and Tennekes 1981) and under some conditions steady state may be reached in a few hours (Brost and Wyngaard 1978, Delage 1974, Surridge 1986a). Theory and observations both support a parabolic growth (Anfossi et al. 1974, Kloppe et al. 1978, Tomasi 1983, Godowitch et al. 1985) of the form

$$h = A t^{1/2}$$

where A is an empirical constant with a value of about $70 \text{ m/h}^{1/2}$ (Anfossi et al. 1974). The stable boundary layer is subject to many complicating influences such as terrain slope, gravity waves (Brost and Wyngaard 1978), advection, surface inhomogeneity, geostrophic wind shear (Zeman 1979) and subsidence (Carlson and Stull 1986) which can alter the heights and rates of growth by 5 to 50%. Equilibrium heights of 200 - 300 m are expected near dawn (Anfossi et al. 1976, Nieuwstadt and Driedonks 1979, Kloppe et al. 1978). Because the growth is dominated by the surface energy balance it is possible to predict the temperature profile from the time series of temperature (Anfossi et al. 1976) or the time series of temperature at two levels (Surridge 1986b). Such a model suggests that the largest temperature difference across an inversion is 63% of the surface temperature drop from the day's afternoon maximum.

The upper air data from Edmonton (Stony Plain) show that the average thickness of surface-based inversions at 05:00 over the years 1971 - 1975 was 241 m with a standard deviation of 126 m. The average intensity of such inversions was 0.023°C/m with a standard deviation of 0.018°C/m . It is commonly observed that the maximum

inversion intensity occurs several hours prior to sunrise. The data for the Edmonton and Calgary towers reflect this. Such warming in the upper levels before sunrise is attributed to subsidence which is also manifest in the lowering of the inversion top (Izumi 1964).

After sunrise on a clear day solar radiation quickly raises the ground temperature. By molecular conduction through a very thin layer the overlying air also becomes warmed. Parcels of heated air tend to move up into the cooler air above while cooler air moves down to replace it. With the presence of a wind the heat transfer is due almost entirely to mechanical turbulence. If the air could adjust instantaneously a layer of constant (potential) temperature equal to that of the ground would be maintained. However, the heating of the air lags far behind the heating of the ground. The temperature gradient across the thin molecular conduction layer is very large and the temperature in the overlying air gradient will be superadiabatic. At greater heights above ground turbulent shear stress becomes weaker and the gradients of wind velocity and temperature diminish. Buoyancy-generated turbulence dominates and the forced convection regime becomes a local free convection regime.

Immediately after sunrise the temperature profile shows a rapid warming in the upper levels, a continuation of the subsidence process that starts before sunrise. Between this upper level warming and the layer affected by surface heating there is an intermediate layer which undergoes a marked cooling (Izumi 1964). This results from the turbulent mixing near the surface which, in destabilizing the surface layer, acts to intensify the overlying inversion and increase the wind shear. Shear-generated turbulence then results in strong mixing which destabilizes this layer. This process is repeated and the mixing propagates upward. The upper portion of the inversion is effectively driven upwards until such time as the mixing is strong enough to eliminate all vestiges of the nocturnal inversion.

With very light winds, a superadiabatic layer near the surface will not form, but rather the inversion near the surface will undergo heating through a relatively deep layer and be destabilized, but not eliminated (DeMarrais 1961). The explanation appears to lie in a process whereby buoyant elements rise and mix with their surroundings while the compensating descent in the environment heats the layer adiabatically. Thus an upward heat flux can be maintained against the temperature gradient.

The evolution of the convective layer can be divided into three phases (Tennekes 1973a). Immediately after sunrise, surface heating acts to "fill" the inversion and change the temperature gradient from stable to unstable. The time for the removal of the inversion is given by

$$t = (2T h_o \Delta\theta / (H_o / \rho c_p))^{1/2}$$

where T is the time scale of the initial heat flux increase (about 3 h), h_o is the initial mixing height, $\Delta\theta$ is the potential temperature difference across the inversion, and H_o is the surface heat flux. If $h_o = 100$ m, $\Delta\theta = 10^\circ\text{C}$, $T = 10$ s, $H_o/\rho c_p = 0.5$ °K m/s (summer), the removal time is about 2 h. In winter with $H_p/\rho c_p = 0.1$ °K m/s, the time of removal of the same inversion increases to around 4 h. During this period the inversion base lifts very slowly and the temperature difference across the inversion (sometimes termed its "strength") decreases.

After the initial inversion has been filled to a point where penetrative convection can begin (generally $h > 3 h_o$), the height of the inversion base begins to rise linearly with time, described by the equation

$$h = (1.4 H_o / \gamma T \rho c_p)^{1/2} t$$

For $\gamma = 0.005^\circ\text{K/m}$ and the other values as above the rise rate will be 425 m/h. Later in the day the rate of growth slows and the height becomes proportional to the square root of time, governed by the equation

$$h = (2.8 H_o / \gamma \rho c_p)^{1/2} t^{1/2}$$

Observations on a 300-m antenna mast near Hamburg, Germany (Kloppel et al. 1978) gave rise rates varying from 48 m/h in September to 84 m/h in April.

Methods purporting to describe boundary layer height in terms of surface properties alone have been criticized because the height and character of the limiting stable layer are products of the history of the airmass in which it is embedded. Thermal winds (Manins 1982), radiation and evaporation will all influence PBL height. Large-scale subsidence can actually cause a PBL to shrink even though heat may be entering from below (Carson 1973).

Another method of estimating the time to eliminate the nocturnal inversion uses eddy diffusivity (Hewson 1945):

$$t = \frac{h^2}{4K}$$

where K is the eddy diffusivity for heat. The suggested value of $K = 3 \text{ m}^2/\text{s}$ appears to be too low to account for the 3-h breakup time of the 400-m inversion in the O'Neill data. The eddy diffusivity may be estimated from the potential temperature gradient (Thomas et al. 1970) as:

$$K = \frac{2.5 \times 10^4}{\rho c_p} \exp(-100\gamma)$$

Values of K observed by the Tennessee Valley Authority ranged from 2 m^2/s to 49 m^2/s .

A third method is based on solar heating (Turner 1967):

$$t = \frac{h^2}{2} \frac{\rho c_p \gamma}{R}$$

where R is the net rate of sensible heating of an air column by solar radiation with a suggested average value of $R = 5.8 \text{ W/m}^2$.

Mathews and Hicks (1978) used acoustic radar records to ascertain that, in Calgary, the nocturnal inversion breaks up $3.0 \pm 0.7 \text{ h}$ after sunrise. Figure 2.8 shows their results. The tower data for Calgary and Edmonton also support this conclusion.

Thomson (1977) used the energy-area proportionality of thermodynamic diagrams to develop an operational method of forecasting mixing heights in northern Alberta. Assuming that the morning temperature gradient is linear and the lapse rate in the mixing layer is near dry adiabatic, the energy required to change the former to the latter is represented by a triangular area on the temperature-height graph. If the available energy due to surface heat flux can be estimated, then the area of the triangle is determined. If the surface temperature increase can be predicted, then the information completely defines the triangle, and the height of the apex (the mixing height) can be calculated. This technique is similar to that of Holzworth but eliminates the need for an observed morning temperature profile. Mixing height categories in 300-m intervals were forecast correctly about 75% of the time.

The hourly values of mixing heights in the Fort McMurray area as reported by Davison et al. (1981) are given in Table 2.9 and were shown earlier in Figure 2.7.

Late in the afternoon, the surface heat flux decreases and the weakening thermals lose their ability to penetrate into the stable layer. The capping inversion acts more and more as a solid barrier. Interfacial mixing decays and the kinetic energy of the thermals is increasingly dissipated rather than converted into potential energy. The downward heat flux decreases accordingly. As solar heating of the ground diminishes, a point of quasi-equilibrium will be reached when ground and air temperatures are the same. With no gradient, conductive heat transfer ceases. Thereafter the temperature of the air will exceed that of the ground and there will be a transfer of heat from the air to the ground. This negative heat flux usually starts an hour or two before sunset. A stable layer forms near the ground and the entire boundary layer will be cooled gradually. The height of the "capping" inversion is lowered and a ground-based inversion builds up. If the surface cooling persists, the upper and lower inversions eventually join, destroying the constant-temperature layer between them. Further cooling then increases the temperature gradient and the mechanisms described at the beginning of this section come into play.

The breakdown of the convective boundary layer is rapid (Caughey and Kaimal 1977); within a matter of minutes an hour before sunset, the level of zero heat flux, normally located near the middle of the convective layer, makes a sudden descent to the surface. The transition to negative heat flux actually propagates downwards to the surface and not upwards as might be expected. There is a lag of about an hour between entrained heat flux and surface heat flux, the maintenance of a negative heat flux in the upper boundary layer when the surface heat flux has dropped to near zero will give the appearance of a downward propagating transition.

Mathews and Hicks (1978) found from SODAR records that the nocturnal inversion at Calgary forms 0.4 ± 0.7 h before sunset. The tower data for Calgary and Edmonton (Choukalos 1985) gave similar results.

The late afternoon behavior including the sudden collapse of the convective boundary layer is most difficult to model. Kumar (1979) used numerical methods

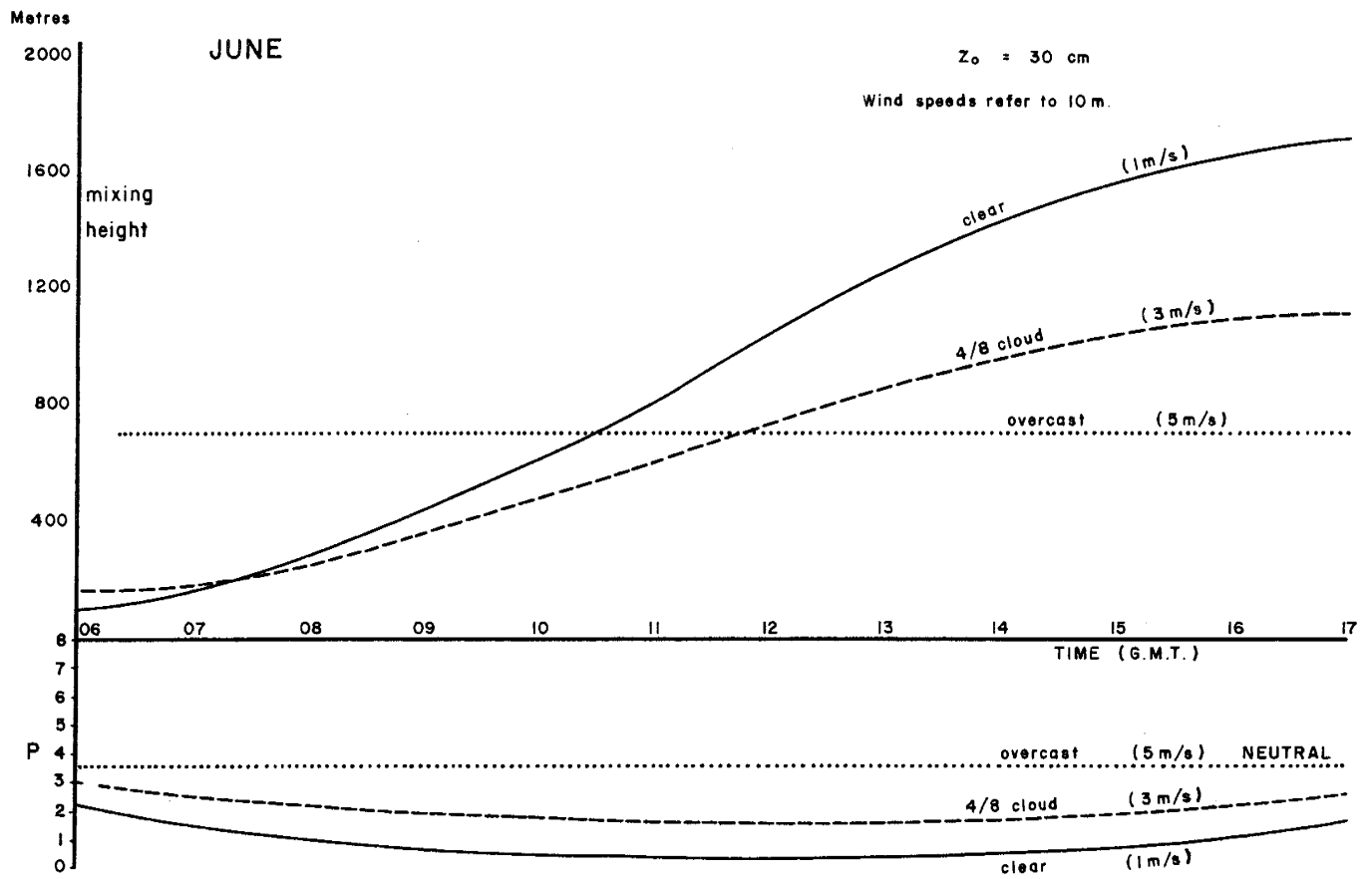
Table 2.9 Seasonal hourly values of mixing height (m) (Davison et al. 1981).

Hour	Winter	Spring	Summer	Autumn
0	0	0	0	0
1	0	0	0	0
2	0	0	0	0
3	0	0	0	0
4	0	70	70	0
5	0	120	120	50
6	0	200	200	100
7	0	250	250	170
8	60	350	350	220
9	120	420	420	350
10	170	500	700	450
11	200	570	850	600
12	220	650	920	770
13	260	750	1000	800
14	270	900	1000	750
15	260	1000	1000	670
16	200	1000	1000	600
17	0	900	900	500
18	0	400	500	0
19	0	0	300	0
20	0	0	50	0
21	0	0	0	0
22	0	0	0	0
23	0	0	0	0

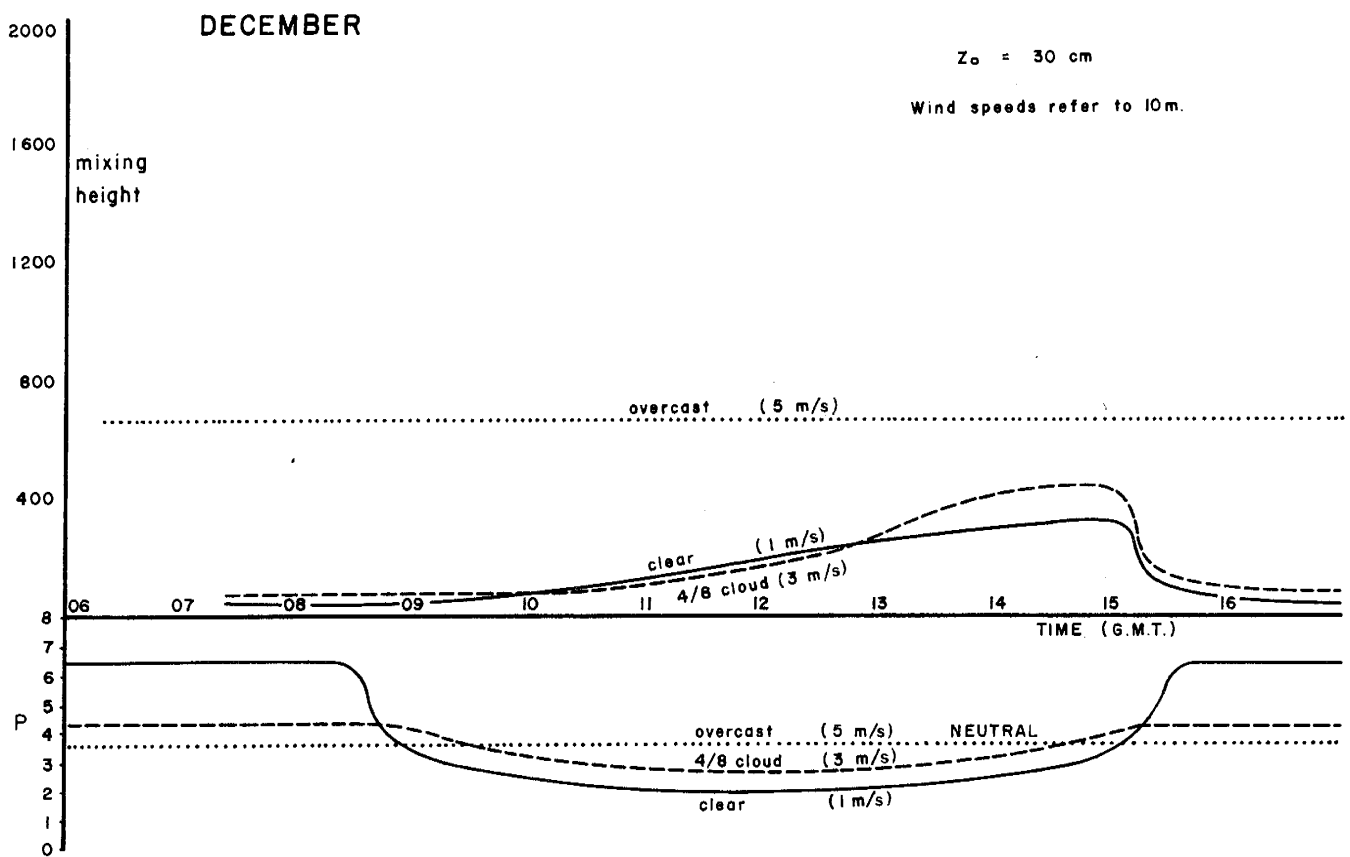
to solve the conservation equation with parameterization for kinetic energy and dissipation rate changes, but was unable to obtain realistic predictions for late afternoon in northeastern Alberta. He attributed the failure to a poor representation of entrainment rates. In contrast, the extended Holzworth method was found to reproduce the sudden collapse of the mixing layer quite well in Florida (Sholtes 1972).

The diurnal variations of mixing height and stability (as represented by Smith's p) for typical days of clear sky, 4/8th cloud and overcast conditions for mid-season in England are shown in Figure 3.9 (Smith 1972). These were based on actual measurements at Cardington and on a developing boundary layer model. They represented mean conditions in steady wind and cloud cover. The effect of cloud cover and/or strong winds is to weaken the amplitude of the diurnal cycle. In overcast windy conditions the boundary layer will have a constant thickness and neutral stability.

The surface heat flux is also the major driving mechanism of mixed-layer growth. Both the height of the convective boundary layer and the nature of the turbulence therein are controlled by the heat flux. Despite its importance it is not routinely measured, largely because of the stringent instrument requirements. Even the most sophisticated sonic anemometer/thermometer eddy correlation system can probably do no better than 20% accuracy, plus it is very difficult to ensure areal representativeness. The usual approach has been one of estimation from more easily measured quantities. Unfortunately the surface heat flux depends on many meteorological variables including cloud cover, albedo, temperature, solar elevation, evaporation, soil moisture and soil heat transfer. Such a formidable task is traditionally handled by discarding all but the most important variables and tolerating the resulting errors. The mean daily global solar radiation estimates of Titus and Truhlar (1969) give an indication of the available energy (Figure 3.10). Many workers including Kumar (1978b) and Thomson (1977) have assumed that the heat flux equals the net radiation, but detailed energy budget parameterizations yield ratios between 0.2 and 0.8 (Wang and Chen 1980). Often it is found that $H = 0.4 R_n$ where R_n is the net radiation (Smith 1979, Misra and Onlock 1982, Tennekes and van Ulden 1974). Because net radiation is not commonly measured, a variety of radiation models have been employed using various astronomical and synoptic weather observations, for example Kumar (1978b), Thomson (1977) and MEP (1977), but standard errors of estimate of 50% are quite common.

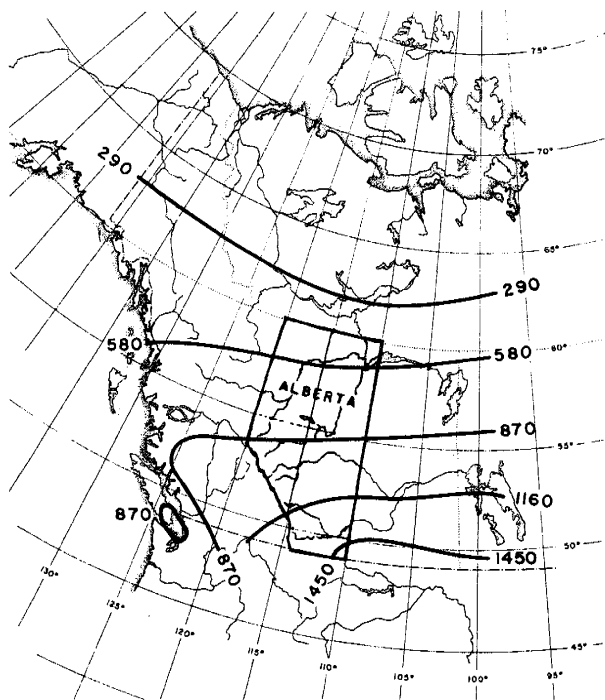


(a) Summer

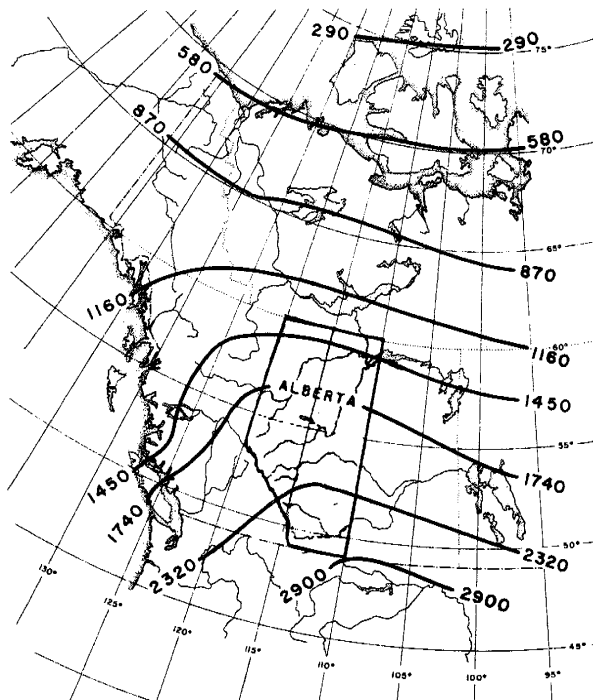


(b) Winter

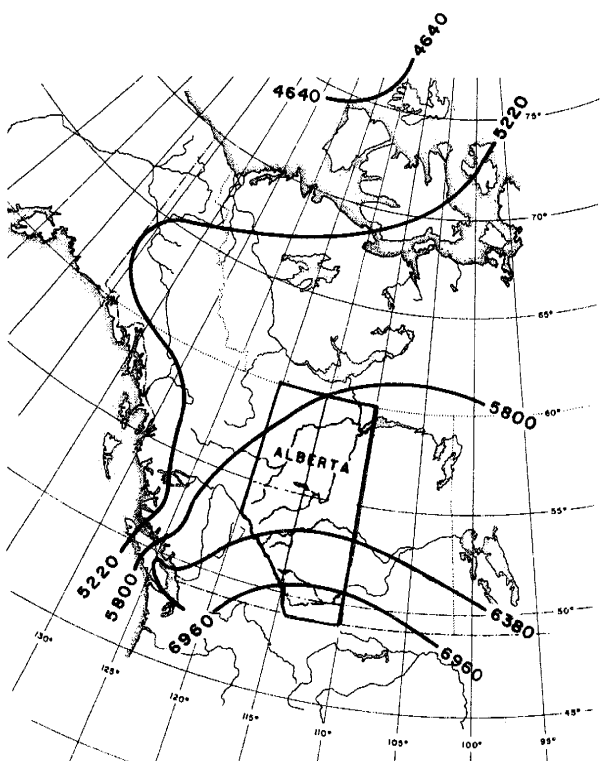
Figure 2.9 Diurnal variation of P stability and mixing heights (Smith 1972).



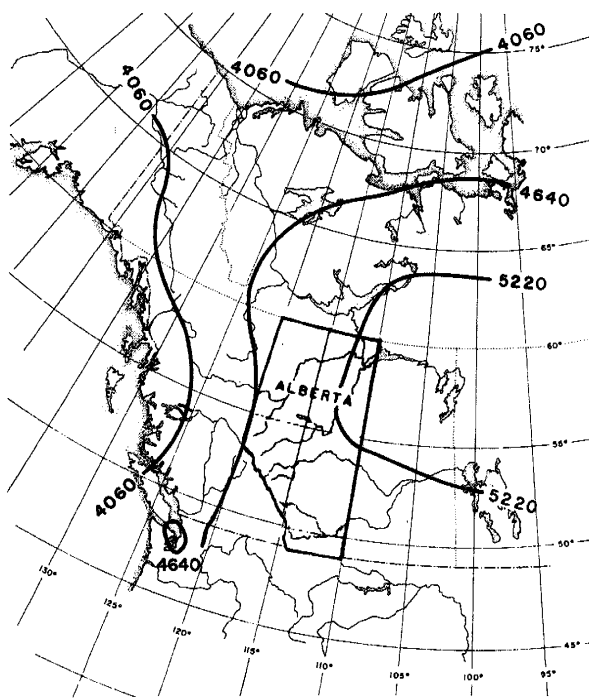
January



October



July



April

Figure 2.10 Mean daily global radiation in Wh/m^2 (Titus and Truhlar 1969).

2.7 PLUME BEHAVIOR

A visible plume can be classified by its appearance (Church 1949) which may resemble any of the shapes given in Figure 2.11. Associated weather conditions and other information are given in Table 2.10.

Coning: Coning is associated with Pasquill stability categories C and D (Gifford 1968, 1975). Leung (1972) found six episodes of high concentration associated with coning in a one-year period for a gas plant northeast of Calgary with two stacks 122 m high and seven monitoring stations.

Fanning: Fanning occurs with temperature inversions, when vertical mixing is suppressed. Crosswind mixing does occur and the plume will spread somewhat in the horizontal. In addition the whole plume may meander, very much as a river does. Concentrations within the plume are high, but little effluent reaches the surface unless high terrain is encountered. Wind shear may enhance the dispersion in the horizontal and plumes at different heights may travel in different directions. If the mean wind direction has shifted over the past few hours, the region may be "tied in ribbons" with portions of the plume from a earlier time still visible.

Looping, Thermalling and Exploding: These three types occur in unstable atmospheres. Looping is the result of eddies which bring the plume to the ground in intermittent "blobs". Within these "blobs" concentrations are likely to be high. Thermalling (Scorer 1958, 1978) results either from neutral updrafts in convective conditions or from updrafts induced by the heat of the plume. Exploding is the rapid vertical spread which occurs when there is strong thermal turbulence confined to smaller sizes than those producing looping.

Considerable controversy surrounds the effect of plume looping on hourly average concentration. Some have argued that at any point the time average of the concentration is low because the wild oscillations cause the plume to be absent from any given point the vast majority of the time, and the average of a few high values with zero values will yield a low concentration (Forsdyke 1970, Turner 1967). Others have argued that since the puffs reach the ground close to the source, average ground-level

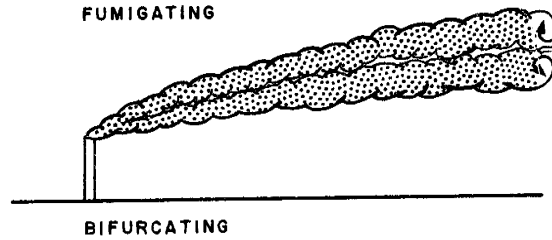
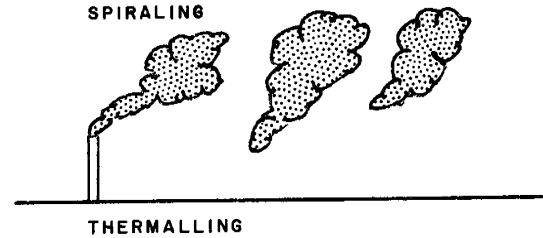
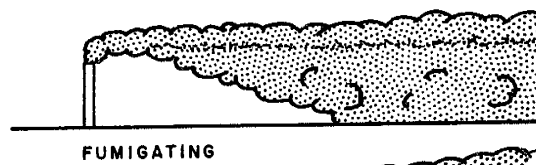
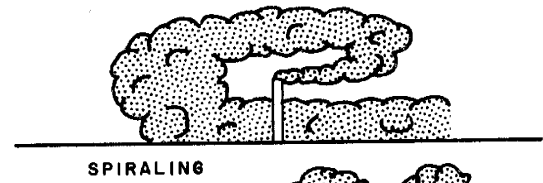
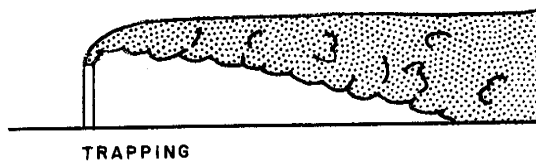
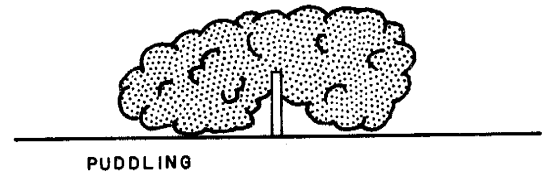
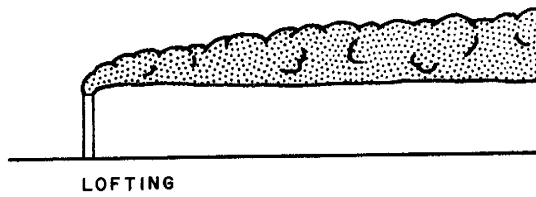
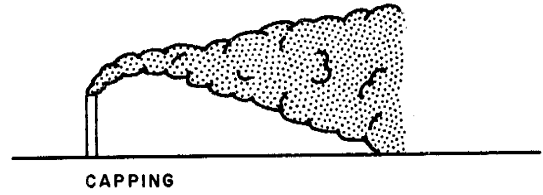
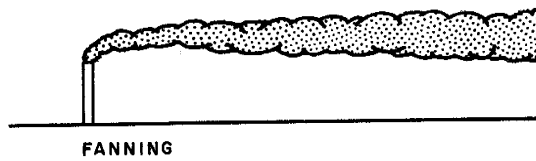
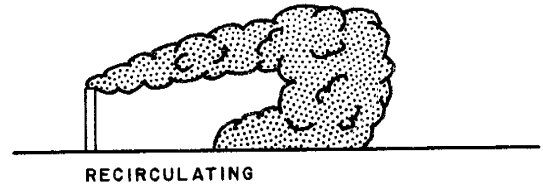
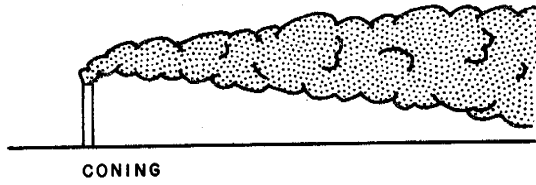
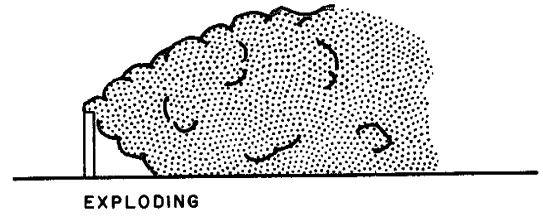
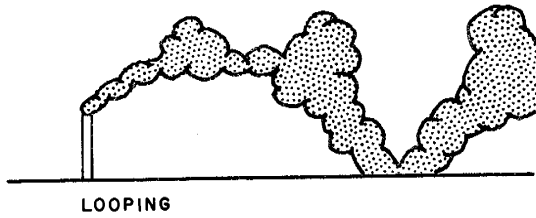


Figure 2.11 Plume behavior types .

Table 2.10 Types of plume behavior and associated meteorology.

Description of Visible Plume	Occurrence	Turbulence/ Stability	Implication for Ground-Level Concentrations
Looping (convective) concentrations irregular loops or waves with apparently random sinuous movements.	daytime with clear skies, light winds and strong solar heating.	strong thermal turbulence (unstable).	high where loop touches ground.
Coning long, slender cone.	strong winds or overcast sky, day or night.	strong mechanical turbulence (neutral) distance.	concentrations rise to a maximum and decrease with distance.
Fanning narrow, horizontal fan widening but not thickening. may resemble a meandering river.	night and early morning, clear skies and light winds. winter days with snow cover and weak sunshine.	mechanical turbulence damped by thermal stratification. (stable).	dilution within the plume is slow, where plume passes near the surface high concentrations will occur.
Lofting coning for some distance, then lower edge, becomes horizontal.	transition from daytime mixing to nocturnal inversion. nighttime, when plume height exceeds inversion depth.	upper layer has residual turbulence from daytime mixing neutral or unstable). lower layer has mechanical turbulence damped by thermal stratification (stable).	plumes in upper layer will produce virtually no ground-level concentrations because the surface based inversion acts as a barrier.

Table 2.10 Types of plume behavior and associated meteorology. (cont'd)

Description of Visible Plume	Occurrence	Turbulence/ Stability	Implication for Ground-Level Concentrations
Fumigating (inversion breakup) upper edge horizontal, lower edge ragged and swirling to ground.	during transition from nocturnal inversion to daytime mixing. • when plume travelling in a stable regime encounters a turbulent internal boundary layer.	upper layer has little turbulence (stable). lower layer has mechanical and thermal turbulence (neutral or unstable).	the sudden downward mixing of a fanning plume leads to high concentrations more or less simultaneously along the full length of the plume.
Bifurcating divides into two rolls.	later afternoon/ early evening with light winds. large plume buoyancy.	weak turbulence (near neutral).	same as coning unless plume rise reduced.
Trapping (limited mixing) cones or loops and at some distance upper edge becomes horizontal.	inversion aloft as a result of a frontal zone between airmasses, anticyclonic subsidence, or convection.	upper layer has little turbulence (stable). lower layer has mechanical and thermal turbulence (neutral or unstable).	after the plume edge contacts the inversion, upward diffusion is prevented, leading to higher concentrations at ground level.
Puddling (calm subsidence) mushrooms above stack or drifts in various directions but never leaves area.	stagnant air near centre or eastern side of an anti-cyclone. weak pressure gradient, winds light and variable.	upper layer has little turbulence (stable). lower layer has weak turbulence (near neutral).	where the edge of the mushroom touches down high concentrations are experienced.

Table 2.10 Types of plume behavior and associated meteorology. (cont'd)

Description of Visible Plume	Occurrence	Turbulence/ Stability	Implication for Ground-Level Concentrations
Recirculating travels away at high level and returns at lower level.	winds at upper levels in opposite direction to those at lower levels.	mechanical turbulence, strong near the interface between the flow zones (near- neutral).	the returning plume can cause high concen- trations near the stack.
Spiraling coils back on itself.	light winds and capping inversion. wind direction changing with altitude or time.	upper layer has little turbulence (stable). lower layer has weak turbulence (neutral or slightly unstable).	because the plume does not leave the area, concentrations can be high.
Capping (wind shear) plume rise abruptly and prematurely terminated.	large, sharp change in wind velocity with altitude.	strong mechanical turbulence in vicinity of wind change (neutral).	reduction in rise and strong vertical mixing produces high concentrations.
Exploding plume expands rapidly in vertical.	light wind and clear sky.	strong thermal turbulence (unstable).	high concen- tration close to stack.
Thermalling plume breaks up in separate masses that rise.	light wind, strong solar heating. large plume buoyancy.	thermal turbulence (unstable).	ground-level concentrations will be very small.

concentrations are likely to be quite high (Slade 1968, Smith 1968). Still others hold the view that although vigorous down currents may produce short-term (3-minute) concentrations as high as any experienced throughout the year, the time duration of these concentrations is small, and the hourly averages will be no larger than for other conditions (Moore 1974, Lucas 1975b). Some workers (Moore 1967) have noted that looping is characteristic of a small plant, with stacks below 100 m and is seldom observed in plants with stacks higher than 150 m; however, others found that with taller stacks looping replaced coning as the condition leading to maximum concentrations (Thomas et al. 1963).

With large heat emissions the plume is rarely brought to ground level by small thermals (Thomas et al. 1963). The large thermals that produce loops of several kilometres for small power plants may provide even greater dilution for larger plants due to the magnitude of the induced circulation. When thermal convection is present the heat emissions from a large plant can initiate thermals and establish a semi-independent local atmospheric circulation. The plume may then rise a few thousand metres and produce cumulus clouds. A heat source of 4 MW over a 500-m square area has been described as adequate for the initiation of thermals (Scorer 1958). Since many Alberta industrial sources produce greater amounts of heat, large, strong thermals should be produced, resulting in a reduction of pollution at ground level. If pollution is carried to the height reached by thermals, typically cloud base height of about 1 km, afterwards being diluted downwards by the ambient turbulence, then the concentration experienced at the ground will be negligible compared with those sources of pollution not producing thermals (Scorer 1958).

Benjamin (1975) examined the consequences of a particular model of convection and concluded that if the plume is emitted into a thermal it will not reach the ground, but if it is emitted into the subsiding environment maximum ground-level concentrations are increased by a factor of 2 or 3 over the coning values. The predictions were not tested against experimental data.

All of the foregoing suggest that it is advantageous for industrial sources to emit into a thermal. If the stack is itself a large heat producer, such a thermal will be self-induced. If the stack is only a small heat source it would be advantageous to locate near a "hot spot" favorable for the formation of thermals. Glider pilots are well aware

of such areas and may provide excellent guidance in this regard. Slopes that face the sun and arid regions commonly act as fixed thermal sources.

Similarly, it is disadvantageous to be located in areas of subsidence such as those found over a river or lake because the plume will be pulled to the ground. However, the frequency of occurrence of high concentrations at ground level due to subsiding air will decrease as the heat emissions increase (Moore 1967).

A looping frequency of 120 h/year has been reported at High Marnham in England (137-m stacks). The ratio of summer to winter looping is a function of latitude and in England it is greater than 3 to 1. At Risø, Denmark (latitude 55°N) looping from a 60-m stack was observed less than 3% of the time in winter but as much as 25% of the time in summer (Frankenburg et al. 1967).

The phenomenon of plume looping is intrinsically tied to the process of convection. Plume rise limited by convection (Briggs 1975) produces ground-level concentrations higher than for the coning case provided the downdraft velocities exceed the buoyancy scale velocity of the stack. This condition can be fulfilled only for sources with small heat emissions. It has also been suggested that diffusion is enhanced under looping conditions because the pulsations of wind speed greatly affect the height of rise of successive plume sections and make a considerable contribution to the time-mean spread (Pasquill 1974, Halliday 1968). Laboratory experiments in a convection chamber (Deardorff 1985) show that the concentration maximum descends downward toward the ground because the downdrafts cover more area than the energetic updrafts. The maximum moves upward thereafter since more and more of the effluent gets swept into updrafts.

Venkatram (1980) applied a convective dispersion model to the Suncor and Syncrude stacks in the Alberta oil sands and predicted ground-level concentrations slightly larger than those for coning, with maximum concentrations occurring between 1500 and 6500 m. Strosher and Peters (1980a, b) identified convective mixing as the cause in two of eight cases of high concentrations. Fanaki et al. (1979) observed that plume looping occurred quite frequently with the Suncor plume near Fort McMurray during the summer season near noon. The plume touched down at distances between a few hundred metres and 1 km, producing high concentrations over the few-minutes sampling period.

Trapping: Trapping occurs when upward diffusion of a plume is restricted by an inversion, below which there is strong mixing. This barrier or "lid" limits the vertical growth of the plume and any decrease in concentration must be due to horizontal spreading.

The capping inversion of the convective boundary layer represents such a lid, but in well developed convection it will be at a height of around 1 km and unlikely to have an effect on most industrial plumes. Moreover, plume looping will be the dominant behavior under such conditions.

Of probably greater significance are large-scale meteorological processes that lead to stabilization aloft. In the stagnation-prone eastern United States, subsidence inversions associated with anticyclones are the most important cases. In Alberta, frontal inversions and airmass inversions are likely to be more significant. Another instance of trapping occurs briefly in conjunction with inversion breakup fumigation. After the nocturnal radiation inversion is eroded past stack height, the plume is emitted into a mixing layer bounded on the top by the remnants of the rising nocturnal inversion. Thus all of the near stack ("younger") sections of the plume experience a limit to upward growth until the far stack ("older") plume sections are fumigated. After the fumigation, the whole plume experiences trapping until the nocturnal inversion is completely destroyed.

With respect to daytime boundary layer evolution, the *potential mixing depth*, represented by the lowest major inversion base can be distinguished from the *effective mixing depth* (Pasquill 1974). Vertical mixing is effective up to the inversion only for the part of the day when surface heating generates sufficiently vigorous convective motion. At other times of the day, and especially in periods of strong nocturnal radiative cooling of the surface, vertical mixing may be confined to a height well below the overhead inversion. Typically there would be diurnal progression of effective mixing depth with a minimum at night and a maximum during the afternoon. As a consequence, at any given time there would be some upwind distance within which the vertical distribution would still be in the process of adjustment, but beyond which vertical spreading would have been completed to the stage of uniformity of concentration with height.

One can also make a distinction between a passive *mixed* layer and an active *mixing* layer (Holzworth 1974). While the mixing layer at some time may be relatively shallow, pollutants at the same time may be found to be mixed through a much deeper layer corresponding to the mixing height at a previous time (McCaldin and Sholtes 1970). When an inversion layer descends, pollutants immediately below the base are left behind and eventually transported and diffused in the atmosphere above the boundary layer. Thus vertical oscillations of an overhead inversion constitute a mechanism for pumping pollution out of the boundary layer. Theory shows that the percentage of pollution pumping in a 24 h period (due to diurnal inversion base ascent and descent) is a function of the ratio of the maximum to minimum inversion base height (Kao 1979). If the inversion base remains stationary, no material is removed. If the maximum is 10 times the minimum, 58% is removed.

Although it is well known that plume trapping leads to ground-level concentrations larger than the coning values, the exact amount of the increase is uncertain. In theory, if the inversion is treated as a completely impervious barrier that totally reflects the plume, with the worst case of an inversion base at plume height, the maximum ground-level concentration is doubled (Scriven 1967, Healey and Baker 1968, Heines and Peters 1973, Ragland 1976). Some measurements confirm this (Leavitt et al. 1971, Moore 1967, Martin and Barber 1973) but others support a threefold increase (Carpenter et al. 1976) which can be predicted by a semi-empirical model for tall stacks up to 305 m (Carpenter et al. 1971). A factor of 3 might be experienced rather than a factor of 2 because the trapping inversion restricts the plume rise (Gifford 1975). Whaley and Lee (1973) indicated smaller increases because the capping inversion causes a widening of the plume. Whaley (1969) proposed that a factor of 2 be used as a Canadian standard with an estimated 12% occurrence for plumes under 200 m.

Leakage through an inversion could have significant implications on ground-level concentrations (Scriven 1967). If, for example, an inversion is assumed to have a diffusivity of only 1% of that in the mixed layer instead of zero, the maximum ground-level concentration is decreased by about 10% in the worst case of inversion base and plume height coinciding. In convective conditions, material beneath an inversion cannot penetrate through it but material above the inversion readily passes through and affects the whole convection layer because of entrainment from aloft (Ball 1960).

When middle tropospheric inversions are associated with mechanical mixing above the inversion, the inversion will migrate downwards and pollutants will be able to pass upwards through the inversion, but not downwards.

A perfectly reflecting barrier above the plume has little or no effect on either the value or position of the maximum ground-level concentration provided it occurs at a vertical distance of 1.25 effective stack heights or more above the ground (Scriven 1967, Heines and Peters 1973). It is when the reflecting barrier (impenetrable inversion) is between 1 and 1.25 effective stack heights above ground that the maximum ground-level concentration is very sensitive to the height of the barrier. With a stable layer aloft, plumes do not achieve their normal plume rise because part of the buoyancy is taken up in lifting and/or destroying the lower part of the stable layer (Martin and Barber 1973). However, a plume would probably not be fully reflected from the stable layer and, if a reflection of 80% is assumed, ground-level concentrations would not exceed a factor of 2 more than the concentrations that would have occurred without the elevated stable layer above. It can be shown that if an elevated stable layer restricts the plume rise to only 75% of the value that would have occurred in the absence of the stable layer, then 50% penetration of the stable layer will occur (Briggs 1975). The net result is that maximum ground-level concentrations would be increased by less than a factor of 2 over the coning case.

In the Tennessee Valley maximum concentrations occurred within 5 to 10 km of a 244-m stack and persisted for 2 to 5 h (Carpenter et al. 1976). High pressure systems leading to limited mixing were found to occur on 30 to 40 days per year.

Leung (1972) reported nine instances in one year of high concentrations associated with plume trapping at a gas plant northeast of Calgary. One of these was associated with the presence of a warm front. Strosher and Peters (1980a, b) found that 5 of 16 episodes in northeastern Alberta were associated with frontal inversions.

Lofting: Lofting occurs with a near-neutral layer above a surface-based inversion that does not extend to plume height. Upward diffusion is good, but downward diffusion is inhibited. The inversion acts as a barrier to the effluent and keeps the ground-level concentration low. Large particulates, however, will still fall though the

inversion by gravitational settling. Depending on the effective height of the stack and the rate of deepening of the inversion, lofting may be very transitory or may last for several hours. It is usually observed for an hour or so at sunset when the lower atmosphere is rapidly changing from unstable/neutral to stable. Such lofting conditions are ideal for the dispersion of industrial stack effluents. Site-specific studies will determine the potential frequencies and whether or not stack design can be modified to take advantage of prolonged lofting.

Fumigating (Inversion Breakup): Inversion breakup fumigation occurs when daytime solar heating has produced a neutral or unstable layer that reaches a plume that was emitted into the nocturnal radiation inversion. The fanning plume contains high concentrations of effluent, and when the mixing layer builds up to plume height, thermal turbulence will bring the effluent to the ground more or less simultaneously along the full length of the plume. This condition gives rise to the greatest ground-level concentrations near a stack over a period of about 30 minutes (Gifford 1968, Turner 1967). For stacks with small plume rise the maximum concentrations occur close to the source but last only a few minutes (Gifford 1975). For stacks with large plume rise the maximum concentration occur several tens of kilometres from the stack and last for 30 to 45 minutes (Gifford 1975). The explanation for this lies in the time taken for the inversion to be eliminated to plume height. Because a fanning plume is a prerequisite for an inversion breakup fumigation, only the portions of the plume that were actually emitted into the inversion layer can be fumigated. These "older" emissions have travelled a considerable distance downwind during the time it takes for the inversion layer between stack height and plume height to be destroyed. The "newer" portions of the plume are emitted into the developing mixing layer and hence cannot be fumigated in the same manner.

For one of TVA's small steam plants the 30-minute average concentration due to inversion breakup was only 0.6 that due to coning (Thomas et al. 1963). For 150-MW units with 76-m stacks it was found that maximum surface concentrations during inversion breakup were approximately equal to those for coning conditions (Thomas et al. 1970). A validated model predicted that for 600- to 700-MW units with stacks of 180+ m the maximum concentrations due to inversion breakup would be 1.5 times those due to coning. The ratio of fumigation to coning concentrations ranged from 1.3 for a 60-m stack to 2.4 for stacks up to 365 m (Carpenter et al. 1971).

Concentrations from British power stations were reported no greater in fumigation conditions than in steady-state conditions with similar wind speeds and lapse rates (Martin and Barber 1973) except for sources in valleys (Lucas 1975a). The maximum concentrations occur within a narrow band at relatively great distances from the source with considerable directional variability (Thomas et al. 1970). In the absence of channelling due to terrain features the frequency of recurrence at the same location is likely to be much less than with coning dispersion (Gifford 1975).

Leung (1972) found six occurrences of inversion breakup episodes in one year at the gas plant northeast of Calgary.

Puddling: Puddling is described as the plume filling an imaginary container around the source (Lott 1985). The velocity vector averaged over long time periods would be zero, indicating that the plume might drift in different directions but would not leave the region. This phenomenon has also been called *calm subsidence* (Gifford 1975), referring to the causal conditions of calm winds in the presence of a subsidence inversion. These meteorological conditions occur near the centre of a stagnating anticyclone (Moore et al. 1986, Holzworth 1969). Such situations appear to be rare, even in the relatively stagnation-prone southeastern U.S. (with an average 5 to 15 stagnation days per year). Only one such occurrence was detected in all the years of TVA monitoring (Briggs 1969). In that instance the ground-level concentration was 50% higher than in the high-wind coning situation and the condition persisted throughout the afternoon. This situation may be the cause of high air pollution in Japan (Ukeguchi et al. 1973). In Alberta the Arctic airmass inversion with calm or light winds can lead to puddling. Strosher and Peters (1980a, b) reported that 3 of 16 high-concentration events in northeastern Alberta were associated with this condition.

Bifurcating: Plume bifurcation has long been observed. It is the result of vorticity which exists at the edges of the stack orifice where there is a large velocity shear between the effluent and the atmosphere (Scorer 1958). The bent-over plume may be imagined to consist of a series of puffs overlapping so that the vorticity at the front of one cancels that at the back of the next. Along the edges of the plume, however, there is no cancellation, so the plume takes the form of two rolls. The downwind extent of this configuration is limited by the growth of the plume since entrainment reduces the

vorticity. Conditions favoring downwind extent are large plume buoyancy, weak thermal stratification and low turbulence. A typical occurrence of these conditions is in the evening, after the thermals of the day's convection have ceased and the radiation inversion is still below the plume.

Bifurcation can also be produced by cooling on the edges of a plume when the plume contains liquid water droplets. Mixing with unsaturated air at the edges of the plume causes some of the droplets to evaporate, extracting the latent heat of evaporation. The plume edges are then cooler than the central part of the plume, a condition which leads to velocity shear and the generation of vorticity.

If the atmosphere is turbulent, bifurcation disappears within a few stack diameters downwind, whereas if the atmosphere is not turbulent, bifurcation may extend for hundreds or thousands of metres downwind (Moore 1967). Coning plumes are usually bifurcated when viewed from beneath rather than from the side (Scorer 1978). Bifurcation may lead to reduced plume rise and higher ground-level concentrations (Fanaki 1975).

Recirculating occurs when the winds at lower levels are opposite to the winds at plume height. *Spiralling* occurs when there are changes in wind direction with altitude or time which cause the plume to coil back on the stack. *Capping* (wind shear) occurs when there is a large discontinuous change in wind velocity with altitude (Lott 1985). At the Paradise steam plant in the Tennessee Valley wind shear capping causes the highest concentrations observed (Lott 1985). Mohtadi and Rowe (1973) reported 8 episodes out of 28 at a gas plant northwest of Calgary could not be attributed to coning, trapping or fumigating and therefore may have been caused by some other atmospheric condition. The frequencies of the various plume behaviors can be roughly estimated from monostatic acoustic sounder records. Mathews and Hicks (1978) reported that at Calgary stable layers (fanning plumes) were observed about 65% of the time, while neutral layers were seen only 7% of the time. Upper layers were present in 33% of the records as shown in Table 2.11.

Table 2.11 Monthly distribution of layer stability from acoustic sounder records at Calgary (Mathews and Hicks 1978).

Month	Total Stable	Total Unstable	Total Neutral	Stable With Upper Layers	Unstable With Upper Layers	Chinooks
March 76	51.9	43.1	5.0	11.6	6.5	0
April	58.3	39.2	2.5	19.9	4.5	0
May	51.6	42.1	6.3	23.9	6.3	0
June	47.6	42.4	9.8	19.0	3.9	0
July	49.1	43.2	7.7	13.5	4.2	0
August	54.2	36.9	8.9	21.1	7.9	0
September	58.7	30.0	11.3	19.1	8.2	0
October	65.0	25.2	9.8	21.5	6.1	0
November	74.9	13.7	11.4	32.6	7.4	2.7
December	89.1	9.6	1.3	40.2	9.6	17.1
January 77	93.4	3.7	2.9	49.7	3.7	3.3
February	81.5	15.4	3.1	40.4	3.4	3.4
March	61.4	33.0	5.5	36.8	4.4	0
Means	64.9	28.5	6.6	27.6	5.8	2.2

2.8 SUMMARY

Mixing heights have been measured at three locations in Alberta. Mean maximum afternoon mixing heights at Edmonton (Stony Plain) range from 242 m in winter to 1901 m in summer. Near Fort McMurray, median maximum afternoon mixing heights range from 270 m in winter to 1000 m in summer. Noontime values at Edmonton (Ellerslie) averaged 162 m in winter and 683 m in summer. Mean maximum afternoon mixing heights have been mapped across Alberta by interpolation from Canadian upper air stations. Northern Alberta has lower mixing heights than southern Alberta; in winter, northern Alberta values are about 200 m while southern Alberta values are about 400 m; in summer, northern Alberta values are about 1600 m while southern Alberta values are about 2200 m. There is a tendency to larger mixing heights in the foothills and mountains, but this may be the result of the interpolation; few mixing height measurements are available for the east slopes in Alberta.

Mean wind profiles have been measured in a variety of Alberta locations, with the results generally presented in terms of the power law. Boundary layer scaling has not been widely applied. Power law exponents vary with surface roughness, stability and layer thickness. On an annual basis, for layers several hundred metres deep, observed power law exponents generally fall between 0.12 and 0.50.

Nocturnal radiation inversions generally break up 3 h after sunrise and form at sunset. Methods for predicting the evolution of the boundary layer has been applied with some success in areas of Alberta with gentle terrain.

Different plume behavior types have different consequences for air quality. The potential frequencies of these types that can produce high ground-level concentrations are: coning 7%, looping 28%, trapping 33%, fumigating 65%. Actual frequencies for any particular plume will be less because the occurrence depends on plume height and buoyancy.

2.9 REFERENCES

- Anfossi, D., P. Bacci and A. Longhetto. 1976. Forecasting of vertical temperature profiles in the atmosphere during nocturnal radiation inversions from air temperature trend at screen height. *Quart. J. Roy. Meteor. Soc.* 102: 173 - 180.
- Anfossi, D., P. Bacci and A. Longhetto. 1974. An application of lidar technique to the study of the nocturnal radiation inversion. *Atmos. Environ.* 8: 537 - 541.
- Angle, R.P. 1975. Airflow from mustard to fallow. *Atmosphere* 13: 110 - 125.
- Arya, S.P.S. 1981. Parameterizing the height of the stable atmosphere boundary layer. *J. Appl. Meteor.* 20: 1192 - 1202.
- Ball, F.K. 1960. Control of inversion height by surface heating. *Quart. J. Roy. Meteor. Soc.* 86: 483 - 494.
- Benjamin, S.F., 1975. The behavior of hot plumes from tall chimneys under conditions of strong thermal convection. Paper presented at the Annual Meeting of the Pacific Northwest International Section, Air Pollution Control Association, Nov. 1975. Vancouver, B.C. 41 pp.
- Benkley, C.W. and L.L. Schulman. 1979. Estimating hourly mixing depths from historical meteorological data. *J. Appl. Meteor.* 18: 772 - 780.
- Bernstein, A.B. 1967. A note on the use of cup anemometers in wind profile experiments. *J. Appl. Meteor.* 6: 280 - 286.
- Blackadar, A.K. 1957. Boundary layer wind maxima and their significance for the growth of nocturnal inversions. *Bull. Amer. Meteor. Soc.* 38: 283 - 290.
- Boers, R. and E.W. Eloranta. 1986. Lidar measurements of the atmospheric entrainment zone and the potential temperature jump across the top of the mixed layer. *Boundary Layer Meteor.* 34: 357 - 375.
- Briggs, G.A. 1975. Plume rise predictions, in lectures on air pollution and environmental impact analyses. Chap. 3: 59 - 111. Boston, Sept. 29 - Oct. 3, Amer. Meteor. Soc., Boston, Mass.
- Briggs, G.A. 1969. Plume rise. US Atomic Energy Commission Critical Review Series. Available as TID 25075 from NTIS. Springfield, VA. 81 pp.
- Brost, R.A. and J.C. Wyngaard. 1978. A model study of the stably stratified planetary boundary layer. *J. Atmos. Sci.*, 35: 1427 - 1440.
- Buckler, S.J. 1969. The vertical wind profile of monthly mean winds over the prairies. Canada Dept. of Transport. Meteorological Branch, TEC 718, June 3, 1969. 16 pp.
- Businger, J.A. and S.P.S. Arya. 1974. Height of the mixed layer in the stably stratified planetary boundary layer. *Advances in Geophysics*, Vol. 18A, Academic Press. 73 - 92.

- Carlson, M.A. and R.B. Stull. 1986. Subsidence in the nocturnal boundary layer. *J. Clim. Appl. Meteor.* 25:1088 - 1099.
- Carpenter, S.B., W.C. Colbaugh, J.M. Leavitt, W.B. Norris and T.L. Montgomery. 1976. A study of limited-layer-mixing dispersion Bull Run Steam Plant. Tennessee Valley Authority, Div. of Environmental Planning, Chattanooga, Tennessee, Sept. 1974. 73 pp.
- Carpenter, S.B., T. L. Montgomery, J.M. Leavitt, W.C. Colbaugh and F.W. Thomas. 1971. Principal plume dispersion models: TVA power plants. *J. Air Poll. Cont. Assoc.* 21:491 - 495.
- Carson, D.J. 1973. The development of a dry inversion-capped convectively unstable boundary layer. *Quart. J. Roy. Meteor. Soc.* 99:450 - 467.
- Caughey, S.J. and S.G. Palmer. 1979. Some aspects of turbulence structure through the depth of the convective boundary layer. *Quart. J. Roy. Met. Soc.* 105:811 - 827.
- Caughey, S.J. and J.C. Kaimal. 1977. Vertical heat flux in the convective boundary layer. *Quart. J. Roy. Meteor. Soc.* 103:811 - 815.
- Chiba, O. and S. Kobayashi. 1984. Turbulent characteristics of the vertical wind velocity near the critical Richardson number. *Memoirs of National Inst. of Polar Research.* 34:62 - 69.
- Choukalos, M. 1985. Summary of data from Tall Meteorological Towers in Calgary and Edmonton. Alberta Environment. March 1985. 89 pp.
- Church, P.E. 1949. Dilution of waste stack gases in the atmosphere. *Ind. Eng. Chem.*, 41(12):2753 - 2756.
- Davenport, A.G. 1960. Rationale for determining design wind velocities. *Proc. Amer. Soc. Civil Eng. J. Struct. Div.* 86(515) part 1: 39-68.
- Davies, M.J.E. 1982. An intercomparison of wind profiles obtained from a Doppler acoustic wind sensor with those obtained from tracking helium filled balloons. PNWIS-APCA Annual Meeting. Vancouver, B. C. November 15 - 17, 1982, 2B-3. to 2B-3.6.
- Davison, D.S. 1974. The translation velocity of convective plumes. *Quart. J. Roy. Meteor. Soc.* 100:572 - 592.
- Davison, D.S., M. Hansen, R.C. Rudolph and M.J.E. Davies. 1981. Airshed management system for the Alberta oil sands. Volume II: meteorological data. Prep. for the Research Management Division by Intera Environmental Consultants Ltd., and Western Research and Development. AOSERP Report 120. 89 pp.
- Deardorff, J.W. 1985. Laboratory experiments on diffusion: The use of convective mixed-layer scaling. *J. Clim. Appl. Meteor.* 24:1143 - 1151.
- Delage, Y. 1974. A numerical study of the nocturnal atmospheric boundary layer. *Quart. J. Roy. Meteor. Soc.* 100: 351 - 364.
- DeMarrais, G.A. 1961. Vertical temperature difference observed over an urban area. *Bull. Amer. Meteor. Soc.* 42: 548 - 554.

- DeMarrais, G.A. 1959. Wind-speed profiles at Brookhaven National Laboratory. *J. Meteor.* 16: 181 - 190.
- Djurfors, S.G. 1971. Analysis of the Calgary wind data. Investigation of air pollution in Calgary and Edmonton. Section B. prep. by Geoscience Research Associates for Alberta Department of Health, Environmental Health Services. 10 pp. plus Appendices.
- Driedonks, A.G.M. 1982. Models and observations of the growth of the atmospheric boundary layer. *Boundary Layer Meteor.* 23: 283 - 306.
- Dubosclard, G. 1980. A comparison between observed and predicted values for the entrainment coefficient in the planetary boundary layer. *Boundary Layer Meteor.* 18: 473 - 483.
- Fanaki, F.H. 1975. Experimental observations of a bifurcated buoyant plume. *Boundary Layer Meteor.* 9: 479 - 495.
- Fanaki, F., R. Hoff, L. Barrie, R. Mickle, M. Lusi, K. Anlauf, A. Gallant, J. Kovalick, F. Froude, J. Markes, J. Arnold, S. Melnichuk, D. Brymer, A. Gaudenzi, A. Moser and D. Bagg. 1979. Air system summer field study in the AOSERP study area, June 1977. prep. for the Alberta Oil Sands Environmental Research Program by Environment Canada, Atmospheric Environment Service. AOSERP Report 68. 248 pp.
- Forsdyke, A.G. 1970. Meteorological factors in air pollution. WMO Technical Note No. 114. WMO No. 274-TP.153. Geneva, Switzerland. 32 pp.
- Frankenburg, T.T., A.C. Stern, E.C. Halliday, D.J. Moore, R.A. Scriven, G. Nonhebel, P.C. Blokker, D.H. Lucas, A.E. Martin, H.H. Macey. 1967. Symposium on chimney plume rise and dispersion. Discussion. *Atmos. Environ.* 1: 425 - 440.
- Gifford, F.A. 1975. Atmospheric dispersion models for environmental pollution applications. Chapter 2, *Lectures on Air Pollution and Environmental Impact Analyses*, American Meteorological Society, Boston, 35 - 58.
- Gifford, F.A. 1968. An outline of theories of diffusion in the lower layers of the atmosphere. Chapter 3 of *Meteorology and Atomic Energy*. D. H. Slade, editor U.S. A.E.C., available from NTIS as TID -24190, 65 - 116.
- Godowitch, J.M., J.K.S. Ching and J.F. Clarke. 1985. Evolution of the nocturnal inversion layer at an urban and nonurban location. *J. Clim. Appl. Meteor.* 24: 791 - 804.
- Halliday, E.C. 1968. Measurements of the rise of hot plumes. *Atmos. Environ.* 2: 509 - 516.
- Hanna, S.R. 1983. Lateral turbulence intensity and plume meandering during stable conditions. *J. Clim. Appl. Meteor.* 22: 1424 - 1430.
- Hanna, S.R., G.A. Briggs and R.P. Hosker, Jr. 1982. Handbook on Atmospheric Diffusion. U.S. Dept. of Energy DOE/TIC-11223 (DE82002045) 102 pp.

- Hanna, S.R. 1981. Diurnal variation of horizontal wind direction fluctuations in complex terrain at Geysers CA. *Boundary Layer Meteor.* 21:207 - 213.
- Hanna, S.R. 1969. The thickness of the planetary boundary layer. *Atmos. Environ.* 3: 519 - 536.
- Healey, T.W. and R.E. Baker. 1968. Radioactive cloud dose calculations. Chapter 7. *Meteorology and Atomic Energy, US AEC*, 301 - 377.
- Heines, T.S. and L.K. Peters. 1973. The effect of a horizontal impervious layer caused by a temperature inversion aloft on the dispersion of pollutants in the atmosphere. *Atmos. Environ.* 7: 39 - 48.
- Hewson, E.W. 1945. The meteorological control of atmospheric pollution by heavy industry. *Quart. J. Roy. Meteor. Soc.* 71: 266 - 282.
- Holtstag, A.A. M. and F.T.M. Nieuwstadt. 1986. Scaling the atmospheric boundary layer. *Boundary Layer Meteor.* 36: 201 - 209.
- Holzworth, G.C. 1974. Climatological data on atmospheric stability in the United States. *Symp. on Atmos. Diff. and Air Poll.* Sept. 9 - 13, Santa Barbara. Amer. Meteor. Soc.
- Holzworth, G.C. 1969. Large-scale weather influences on community air pollution potential in the United States. *J. Air Poll. Contr. Assoc.* 19: 248-254.
- Holzworth, G.C. 1967. Mixing depths, wind speeds and air pollution potential selected locations in the United States, *J. Appl. Meteor.* 6:1039 - 1084.
- Irwin, J.S. 1979. A theoretical variation of the wind profile power-law exponent as a function of surface roughness and stability. *Atmos. Environ.* 13: 191 - 194.
- Izumi, Y. 1964. The evaluation of temperature and velocity profiles during breakdown of a nocturnal inversion and a low-level jet. *J. Appl. Meteor.* 3: 70 - 82.
- Izumi, Y. and H.A. Brown. 1966. Temperature, humidity, and wind variations during dissipation of a low-level jet. *J. Appl. Meteor.* 5: 36 - 42.
- Jarvis, E.C. 1968a. Objective guidance in forecasting surface winds over land. Canada Dept. of Transport, Meteorological Branch. TEC 690, October 11, 1968. 22 pp.
- Jarvis, E.C. 1968b. An operational method for predicting the height of the planetary boundary layer in a stable atmosphere. Dept. of Transport, Meteorological Branch. Canadian Meteorological Research Reports CMRR 5/68, 18 July 1968. 11 pp.
- Johnson, O. 1959. An examination of the vertical wind profile in the lowest layers of the atmosphere. (Suffield Tech. Paper No. 135, DRB Project D52-32-01-04). *J. Meteor.* 16: 144 - 148.

- Kaimal, J.C., N.L. Abshire, R.B. Chadwick, M.T. Decker, W.H. Hooke, R.A. Kropfli, W.D. Neff and F. Pasqualucci. 1982. Estimating the depth of the daytime convective boundary layer. *J. Appl. Meteor.* 21: 1123 - 1129.
- Kaimal, J.C., J.C. Wyngaard, D.A. Haugen, O.R. Coté, Y. Izumi, S.J. Caughey and C.J. Readings. 1976. Turbulence structure in the convective boundary layer. *J. Atmos. Sci.* 33: 2152 - 2169.
- Kao, S.K. 1979. Oscillation of inversion layer and pollution pumping. 4th Symp. on Turbulence, Diff. and Air Pollution. Jan. 15 - 18, 1979, Reno, Nevada. 448 - 449.
- Klein, W.H. 1957. Principal tracks and mean frequencies of cyclones and anticyclones in the northern hemisphere. Res. Paper No. 40. Weather Bureau. U.S. Dept. of Commerce. 60 pp.
- Klöppel, M., G. Stilke and C. Wamser. 1978. Experimental investigations into variations of ground-based inversions and comparisons with results of simple boundary layer models. *Boundary Layer Meteor.* 15: 135 - 145.
- Kondo, J., O. Kanechika and N. Yasuda. 1978. Heat and momentum transfers under strong stability in the atmospheric surface layer. *J. Atmos. Sci.* 35: 1012 - 1021.
- Kumar, A. 1979. Numerical modeling of the thermal boundary layer near a synthetic crude oil plant. *J. Air Poll. Cont. Assoc.* 29: 827 - 832.
- Kumar, A. 1978a. Pollutant dispersion in the planetary boundary layer. Syncrude Canada Ltd. Professional Paper. 1978-1, 182 pp.
- Kumar, A. 1978b. A simple radiation model for the tar sands area. Syncrude Canada Professional Paper 1978-3, 28 pp.
- Kumar, A. 1978c. Numerical modeling of the development of the thermal boundary layer near a synthetic crude oil plant. Syncrude Canada Ltd. Professional Paper 1978-4. 32 pp. (Presented at 1978 Technical Meeting of Pacific Northwest International Section - Air Pollution Control Association. Paper No. PNWIS 78 - 23. November 8 - 10, 1978. Portland, Oregon).
- Leahey, D. M., M. C. Hansen and M. B. Schroeder. 1988. An analysis of wind fluctuation statistics collected under stable atmospheric conditions at three sites in Alberta, Canada. *J. Appl. Meteor.* 27: 774 - 777.
- Leahey, D. M. and M. C. Hansen. 1985. Doppler acoustic radar observations of wind in the Ellerslie region of Alberta. Prep. for Alberta Environment, Research Management Division by Western Research Division of Bow Valley Resources Services Ltd. RMD Report 85/31. 77 pp.
- Leavitt, J.M., S.B. Carpenter, J.P. Blackwell and T.L. Montgomery. 1971. Meteorological program for limiting power plant stack emissions. *J. Air Poll. Cont. Assoc.* 21: 400 - 405.
- Lee, G.K., H. Whaley and J.G. Gainer. 1973. Plume dispersion at sulphur plants. Energy Processing Canada. Nov. - Dec. 1973, 24 - 34.

- Lester, P.F. 1978. A severe Chinook windstorm. Presented at the Conf. on Sierra Nevada Meteorology. 19-21 June 1978. Lake Tahoe, CA. 14 pp.
- Lettau, H. 1969. Note on aerodynamic roughness-parameter estimation on the basis of roughness-element description. *J. Appl. Meteor.* 8: 828 - 832.
- Leung, T.B. 1972. Sulphur Plant Stack Dispersion. M. Sc. Thesis. Dept. of Chemical Engineering, The University of Calgary, July 1972. 83 pp.
- Leung, T., M.F. Mohtadi and R.D. Rowe. 1972. Evaluation of stack gas sizing formula. *Gas Processing Canada*, May/June 1970, 40 - 42.
- Lott, R.A. 1985. SO₂ concentrations near tall stacks. *Atmos. Environ.* 19: 1589 - 1599.
- Lucas, D.H. 1975a. The effect of emission height in very large areas of emission. *Atmos. Environ.* 9: 607 - 622.
- Lucas, D.H. 1975b. Reply to Discussion. The effect of emission height in very large areas of emission. *Atmos. Environ.* 9: 693 - 695.
- Mahrt, L., R.C. Heald, D.H. Lenschow, B.B. Stankov and I.B. Troen. 1979. An observational study of the structure of the nocturnal boundary layer. *Boundary Layer Meteor.* 17: 247 - 264.
- Manins, P.C. 1982. The daytime planetary boundary layer: a new interpretation of Wangara data. *Quart. J. Roy. Meteor. Soc.* 108: 689 - 705.
- Manins, P.C. 1979. Partial penetration of an elevated inversion layer by chimney plumes. *Atmos. Environ.* 13: 733 - 741.
- Martin, A. and F.R. Barber. 1973. Further measurements around modern power stations. I - III. *Atmos. Environ.* 7: 17 - 37.
- Martin, A. and F.R. Barber. 1967. Sulphur dioxide concentrations measured at various distances from a modern power station. *Atmos. Environ.* 1: 655 - 677.
- Martin, A. and F.R. Barber. 1966. Investigations of sulphur dioxide pollution around a modern power station. *J. Int. Fuel.* 39: 294 - 307.
- Mathews, T. and R. B. Hicks. 1978. A study of the stability of the atmospheric boundary layer as revealed by an acoustic sounder and its relation to air pollution levels in Calgary. prep. for Alberta Environment by Department of Physics, University of Calgary. 87 pp.
- McCaldin, R. O. and R. S. Sholtes. 1970. Mixing height determinations by means of an instrumented aircraft. Paper No. ME-39G. Proc. of the 2nd Int'l Clean Air Congress, Washington, D.C. December 6 - 11. 1180 - 1185.
- MEP Company. 1977. The development of the thermal boundary layer at Mildred Lake, Alberta. prep. for Syncrude Canada Ltd., February 1977. 39 pp.
- Misra, P.K. and S. Onlock. 1982. Modelling continuous fumigation of Nanticoke generating station plume. *Atmos. Environ.* 16: 479 - 489.

- Mohtadi, M. F. and R. D. Rowe. 1973. Dispersion of plumes from sulphur plants. Gas Processing Canada May - June 1973, 16 - 20.
- Monin, A. S. and A. M. Yaglom. 1971. Statistical Fluid Mechanics, MIT Press, Cambridge Mass. 769 pp.
- Moore, D.J., P. Bacci, F. Esteban, P. Lejeune, A. Saab, R.A. Scriven and R. Steenkist. 1986. Summary report of the CORECH (COGAR) workshop on the evaluation of air pollution episodes and associated control measures. Atmos. Environ. 20:2047 - 2052.
- Moore, D. J. 1974. The prediction of the mean hourly average maximum ground level concentration of sulphur dioxide at Tilbury. Atmos. Environ. 8:543 - 554.
- Moore, D. J. 1967. SO₂ concentration measurements near Tilbury power station. Atmos. Environ. 1:389 - 410.
- Murray, W. A. 1980. New approaches to stack design. Energy Processing Canada. Jan - Feb 1980. 48 - 49.
- Murray, W. and M. Morrow. 1977. The vertical wind profile at Mildred Lake, Alberta. Syncrude Canada Ltd. Professional Paper 1977-1. 61 pp.
- Niemeyer, L.E. 1960. Forecasting air pollution potential. Mon. Wea. Rev. 88:88 - 96.
- Nieuwstadt, F.T.M. 1984a. The turbulent structure of the stable, nocturnal boundary layer. J. Atmos. Sci. 41: 2202 - 2216.
- Nieuwstadt, F.T.M. 1984b. Some aspects of the turbulent stable boundary layer. Boundary Layer Meteor. 30: 31 - 55.
- Nieuwstadt, F.T.M. 1980. A rate equation for the inversion height in a nocturnal boundary-layer. J. Appl. Met. 19: 1445 - 1447.
- Nieuwstadt, F.T.M. and H. Tennekes. 1981. A rate equation for the nocturnal boundary layer height. J. Atmos. Sci. 38: 1418 - 1428.
- Nieuwstadt, F.T.M. and A.G.M. Driedonks. 1979. The nocturnal boundary layer: a case study compared with model calculations. J. Appl. Meteor. 18: 1397 - 1405.
- Oke, T.R. 1970. Turbulent transport near the ground in stable conditions. J. Appl. Meteor. 9: 778 - 786.
- Panofsky, H.A. 1974. The atmospheric boundary layer below 150 m. Ann. Rev. Fluid Mech. 6: 147 - 177.
- Panofsky, H.A. and J.A. Dutton. 1984. Atmospheric Turbulence. Models and Methods for Engineering Applications. John Wiley & Sons, New York. 397 pp.
- Panofsky, H.A., A.K. Blackadar and G.E. McVehil. 1960. The diabatic wind profile. Quart. J. Roy. Meteor. Soc. 86: 390 - 398.

- Pasquill, F. 1976. The "Gaussian Plume" model with limited vertical mixing. EPA-600/4-76-042. Aug. 1976. Environmental Sciences Res. Lab. Research Triangle Park, N.C. 20 pp.
- Pasquill, F. and F.B. Smith. 1983. Atmospheric Diffusion. Third edition. Ellis Horwood Limited. Chichester. 437 pp.
- Pasquill, F. 1974. Atmospheric diffusion, 2nd edition. Halsted Press. John Wiley & Sons, New York. 429 pp.
- Pettitt, R.B. and R.G. Root. 1965. Vertical wind shear in the boundary layer. Canada Dept. of Transport, Meteorological Branch. CIR-4345, TEC 590. 9 Dec. 65. 35 pp.
- Plate, E.J. 1971. Aerodynamic characteristics of atmospheric boundary layers. AEC critical review series. US Atomic Energy Commission. 190 pp.
- Pooler, F., Jr. and L.E. Niemeyer. 1971. Dispersion from tall stacks: an evaluation. 2nd Int. Clean Air Congress. Dec. 6 - 11, 1970. Washington, Academic Press, 1049 - 1056.
- Portelli, R.V. 1977a. Joint frequency distribution table of mixing depth parameters. Available from Meteorological Applications Branch, Atmospheric Environment Service, 4905 Dufferin St. Downsview, Ontario. 118 pp.
- Portelli, R.V. 1977b. Mixing heights, wind speeds and ventilation coefficients for Canada. Climatological Studies No. 31. Atmospheric Environment Service, Environment Canada. Available from Supply and Services Canada, Hull, Que. Catalogue No. EN 57-7/31. 87 pp.
- Ragland, K.W. 1976. Worst case ambient air concentrations from point sources using the gaussian plume model. Atmos. Environ. 10:371 - 374.
- Rudolph, R.C. and D.S. Davison. 1985. Estimates of roughness length from minisonde profiles in the Athabasca Oil Sands area. Prep. for Alberta Environment, REsearch Management Division by Intera Technologies Ltd. RMD Report L-87. 42 pp.
- Sakiyama, S. and J. Wong. 1987. Summary of data from the Ellerslie meteorological Station 1978 - 1985. Alberta Environment. October 1987.
- Scorer, R.S. 1974. Seminar on environmental meteorology. University of Calgary, May 23, 1974. 28 pp.
- Scorer, R.S. 1968. Air pollution. Pergamon Press, Oxford. 151 pp.
- Scorer, R.S. 1958. Natural aerodynamics. Pergamon Press. London. 312 pp.
- Scriven, R.A. 1967. Properties of the maximum ground-level concentration from an elevated source. Atmos. Environ. 1:411 - 419.
- Seale, D. 1964. The diurnal variation of wind in the lowest 500 feet. Canada Dept. of Transport, Meteorological Branch, C1R-4126, TEC-541 13 Oct. 64. 4 pp plus maps.

- Senshu, T., S. Yoshida and K. Nakamura. 1974. A model of the boundary layer for the prediction of atmospheric pollution. Symp. Atmospheric Diffusion and Air Pollution, Sept. 9 - 13, 1974, Santa Barbara. American Meteorological Society. Boston. 180 - 183.
- Shaw, R.W. and R.E. Munn. 1971. Air pollution meteorology. Introduction to the Scientific Study of Atmospheric Pollution. B. M. McCormac (ed), D. Reidel Publishing Company, Dordrecht, Holland, 53 - 96
- Sholtes, R.S. 1972. The growth and decay of turbulent mixing in the planetary boundary layer. Florida University, Gainesville. EPA-R4-72-001. 100 pp.
- Slade, D.H. (editor). 1968. Meteorology and Atomic Energy. U. S. Atomic Energy Commission Office of Information Services, TID-24190 from NTIS. 445 pp.
- Smith, F.B. 1972. A scheme for estimating the vertical dispersion of plumes from a source near ground level. Proc. Third Meeting Expert Panel on Air Pollution Modelling. NATO Committee on Challenges of Modern Society. XVII - 1 to XVII - 14.
- Smith, M.E. (editor). 1968. Recommended Guide for the Prediction of the Dispersion of Airborne Effluents. The American Society of Mechanical Engineers, New York. 85 pp.
- Strosher, M.M. and R. Peters. 1980a. Meteorological factors causing SO₂ events from a point source with implications for atmospheric modelling. 2nd Joint Conf. Applications of Air Pollution Meteorology. March 24 - 27, 1980. New Orleans, La. Amer. Meteor. Soc., Boston, 760 - 763.
- Strosher, M.M. and R.R. Peters. 1980b. Meteorological factors affecting ambient SO₂ concentrations near an oil sands extraction plant. prep. for the Alberta Oil Sands Environmental Research Program by Alberta Environment, Pollution Control Division and AOSERP. AOSERP Report 106. 87 pp.
- Surridge, A.D. 1986a. The evolution of the nocturnal temperature inversion. Boundary Layer Meteor. 36:295 - 305.
- Surridge, A.D. 1986b. Extrapolation of the nocturnal temperature inversion from ground-based measurements. Atmos. Environ. 20:803 - 806.
- Tennekes, H. 1973. A model for the dynamics of the inversion above a convective boundary layer. J. Atmos. Sci. 30: 558 - 567.
- Tennekes, H. 1973b. Similarity laws and scale relations in planetary boundary layers. Chap. 5. Workshop on Micrometeorology. D. A. Haugen, ed. Amer. Meteor. Soc. 177 - 216.
- Tennekes, H. and A.P. van Ulden. 1974. Short-term forecasts of temperature and mixing height on sunny days. Symp. on Atmos. Diff. Air Poll., Sept. 9 - 13, 1974. Santa Barbara, CA. Amer. Meteor. Soc., Boston. 35-40
- Thomas, F.W., S.B. Carpenter, J.M. Leavitt, T.L. Montgomery and W.C. Colbaugh. 1970. Report on full-scale study of inversion breakup at large power plants. Tennessee Valley Authority. Muscle Shoals, Alabama, March 1970.

- Thomas, F.W., S.B. Carpenter and F. E. Gartrell. 1963. Stacks - How High? J. Air Poll. Cont. Assoc. 13:198 - 204.
- Thomson, R.B. 1977. An Operational Mixing Depth Model for Northern Alberta. M. Sc. Thesis. Meteorology Division, Dept. of Geography, University of Alberta, Edmonton. 150 pp.
- Thorpe, A.J. and T.H. Guymmer. 1977. The nocturnal jet. Quart. J. Roy. Meteor. Soc. 103: 633 - 653.
- Thyer, N. 1981. Diurnal variations of upper winds in summer in Alberta. Atmosphere Ocean 19: 337 - 344.
- Titus, R.L. and E.J. Truhlar. 1969. A new estimate of average global solar radiation in Canada. Canada Dept. of Transport, Meteorological Branch, CLI 7-69, 3 July 69. 24 pp.
- Tomasi, C. 1983. The nocturnal inversion height in the Po valley. Atmos. Environ. 17: 1123 - 1129.
- Turner, D.B. 1967. Workbook of Atmospheric Dispersion Estimates. Public Health Service. Publication 999-AP-26. Robert A. Toft Sanitary Engineering Centre, Cincinnati, Ohio. 85 pp.
- Ukeguchi, N., H. Okamoto and Y. Ide. 1973. Predictions of calm- inversion pollution. Proceedings of the 3rd International Clean Air Congress. Oct. 8-12. Dusseldorf, B66-B68.
- Van Ulden, A.P. and A.A.M. Holtslag. 1985. Estimation of atmospheric boundary layer parameters for diffusion applications. J. Clim. & Appl. Meteor. 24:1196 - 1207.
- Venkatram, A. 1980. Evaluation of the effects of convection on plume behavior in the AOSERP study area. prep. for the Alberta Oil Sands Environmental Research Program. AOSERP Report 95. 75 pp.
- Venkatram, A. 1978. Estimating the convective velocity scale for diffusion applications. Boundary Layer Meteor. 15:447 - 452.
- Walker, E.R., O. Johnson and W.H. Boswell. 1968. Meteorological characteristics of chinooks in Alberta. Suffield Technical Note No. 227. Dept. of National Defense. 30 pp.
- Wang, I.T. and P.C. Chen. 1980. Estimations of heat and momentum fluxes near the ground. Second Joint Conf. on Appl. of Air Poll. Meteor. March 24 - 27, 1980. New Orleans, LA. Amer. Meteor. Soc., Boston. 764-769.
- Whaley, H. 1969. A proposed Canadian standard for estimating atmospheric dispersion of combustion source pollution from chimneys. Atmos. Environ. 3:177 - 195.
- Whaley, H. and G.K. Lee. 1973. Discussion. The effect of a horizontal impervious layer caused by a temperature inversion aloft on the dispersion of pollutants in the atmosphere. Atmos. Environ. 7:1010.

- Wilson, H.P. 1959. The influence of stability on surface wind behavior. Dept. of Transport, Meteorological Branch. CIR 3266, TEC 312. 12 Nov. 59. 6 pp.
- Yamada, T. 1979. Prediction of the nocturnal surface inversion height. J. Appl. Meteor. 18:526 - 531.
- Yamada, T. and S. Berman. 1979. A critical evaluation of a simple mixed-layer model with penetrative convection. J. Appl. Meteor. 18:781 - 786.
- Yamada, T. and G. Mellor. 1975. A simulation of the Wangara atmospheric boundary layer data. J. Atmos. Sci. 32:2309 - 2329.
- Young, J.W.S. and E. Wilson. 1982. Climatological wind speed profile relationships in Canada. Proc. Fifth International Clean Air Congress, Buenos Aires, October 1980, 679 - 685.
- Yu, T.W. 1978. Determining height of the nocturnal boundary layer. J. Appl. Meteor. 17: 28 - 33.
- Zeman, O. 1979. Parameterization of the dynamics of stable boundary layers and nocturnal jets. J. Atmos. Sci. 36:792 - 804.
- Zilitinkevich, S.S. 1972. On the determination of the height of the Ekman boundary layer. Boundary Layer Meteor. 3:141 - 145.

3.0 PLUME RISE

The height of a chemical plume above the ground comprises of the height at which the release occurred (stack height) plus the rise of the plume above the release point due to buoyancy and momentum effects. For large industrial sources, this total height (effective stack height) can be a factor of 2 to 10 times the actual release height. Because maximum ground-level concentration is approximately proportional to the inverse square of the effective stack height, plume rise can change the maximum ground-level concentration by as much as 100 times.

Plume rise has received a great deal of study by a number of researchers (Sutton 1947, Bosanquet 1957, Scorer 1959, Lucas et al. 1963, Moore 1965, 1984, Slawson and Csanady 1967, Briggs 1975, Shatzmann 1979). Early plume rise models (see Moses and Strom 1961 for a review) were either completely empirical (Holland 1953, Davidson 1949) or semi-empirical (Batchelor 1954, Scorer 1959). There are now areas of agreement among recent models of plume rise (see Briggs 1969), although there are still major gaps in our knowledge.

3.1 GENERAL CONSIDERATIONS

A plume will rise in the atmosphere due to its initial momentum and buoyancy. As the plume rises, a shear zone occurs between the plume and surrounding fluid due to the relative motion of the plume to the ambient air. This shearing results in entrainment of ambient air into the plume (thus reducing the plume's initial momentum and buoyancy) and causing the plume to grow visibly. This "self-generated" turbulence dominates the mixing process close to the source. This initial, self generated turbulence zone is shown in Figure 3.1. Various terms are also defined in the figure. As the plume continues to grow and lose its initial buoyancy and momentum, the atmospheric conditions (wind speed, vertical density stratification, turbulence) begin to dominate the behavior. Equations describing the behavior of the plume (trajectory and spread) during its initial rise phase can be developed by examining the basic laws of fluid mechanics: conservation of mass, buoyancy and momentum. The distribution of temperature, speed, or other quantities across the plume are assumed to have "top hat" form; that is, a variable has a certain value inside the plume, another value outside the plume, and a discontinuity

at the plume radius (R_p). In mathematical terms, the variables are integrated averages in a circular plume cross section. A schematic diagram illustrating the basic definitions for describing plume rise is shown in Figure 3.2.

Because the details of turbulence are neglected, there is one less conservation equation than independent variables; thus a closure assumption must be made to make the equations tractable.

A simple closure assumption (Taylor, 1948 and later Morton, Taylor and Turner, 1956) uses the *entrainment velocity*, v_e , which is defined as:

$$v_e = \alpha, \beta |\bar{w}|$$

where

α, β are the entrainment constants for a jet and bent over plume respectively and $\bar{w}$ is the plume vertical velocity

The entrainment velocity is simply a measure of the entrainment of ambient fluid into the plume caused by the turbulence generated by the velocity of the plume relative to the surrounding air. If the plume is rising in a calm atmosphere, the relative velocity component is simply the vertical velocity of the plume itself, w . This provides the basis for the above equation. For a plume rising in a crosswind, this equation can still be applied assuming that the downwind component of the plume velocity is equal to the windspeed, and the vertical velocity of the plume is much less than the plume velocity (plume is well bent-over).

With further simplifications, the conservation equations (with help from the closure assumption) can now be solved and analytical solutions obtained. The additional simplifications include the Boussinesq approximation ($\rho_a = \rho_p = \text{constant}$, except in the density difference terms), the bent-over plume assumption and the uniform windspeed assumption (where $U(z) = \text{constant}$). Drag effects have been included in some derivations; however, drag has traditionally been ignored by modellers.

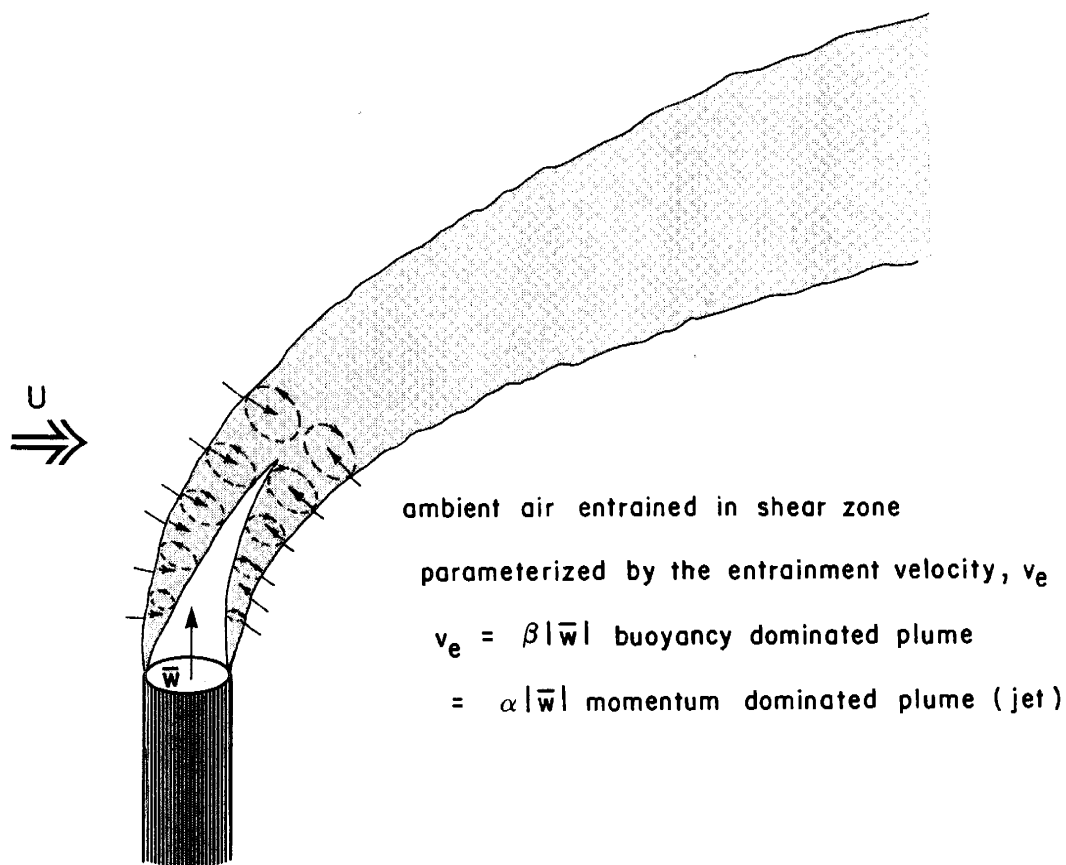


Figure 3.1 Depiction of turbulent shear zone and entrainment.

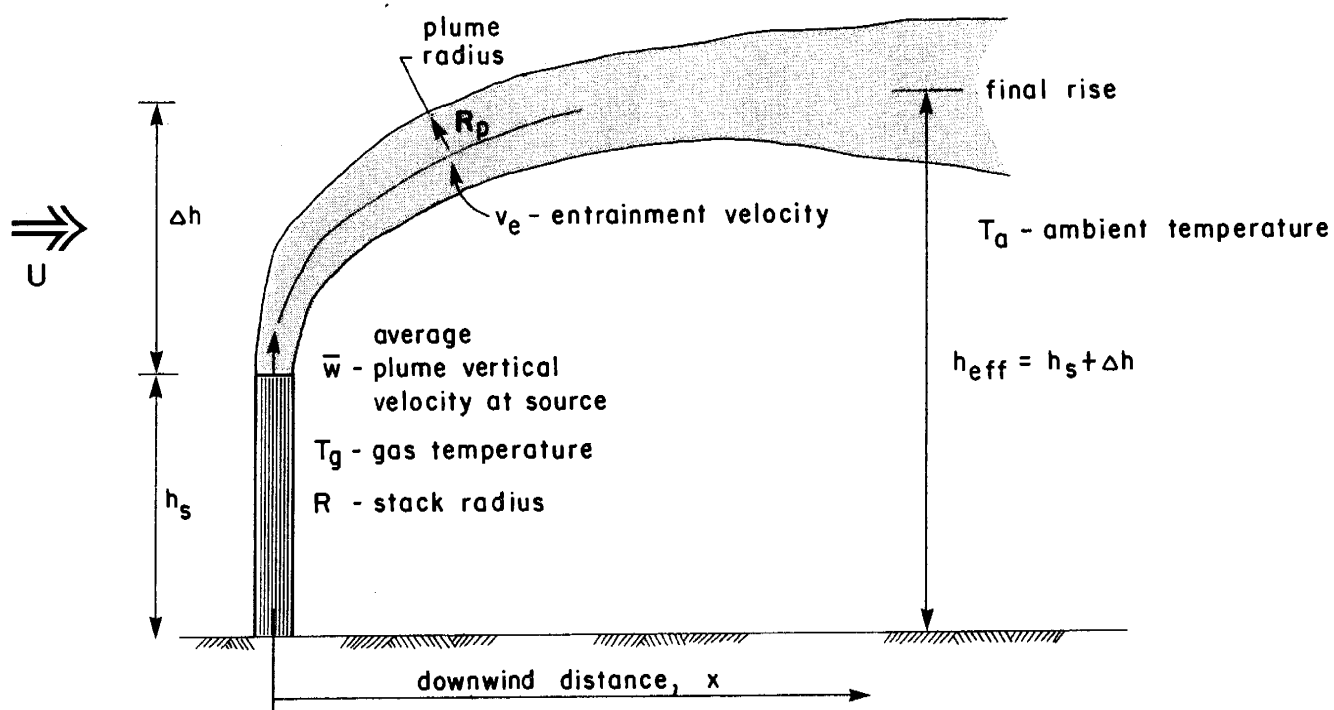


Figure 3.2 Schematic illustrating basic definitions for plume rise.

Djurfors (1977a) solved the conservation equations for mass, vertical momentum and energy for buoyant plumes in calm atmospheric conditions and included drag terms. Drag was characterized by the inclusion of the *drag coefficient*, C_D . The separate effects of entrainment and drag on plume behavior were examined. Djurfors concluded that drag influences plume motion, but has no effect on plume temperatures. In addition, the effect of entrainment on plume motion is the same as drag, i.e., to retard the motion. Even though these results are intuitive, Djurfors illustrated their relative effects mathematically.

Numerical solutions are possible without making the above assumptions. Slawson et al. (1980) used a variety of numerical and analytical plume rise solutions along with a number of different entrainment hypotheses to compare with observations of the Suncor (GCOS) plume. The numerical models performed better than the analytical forms. This would be expected since non-uniform wind and temperature profiles could be precisely accounted for and the Boussinesq and bent-over plume assumptions could be relaxed. The performance of current regulatory plume rise algorithms and a new numerical plume rise model were compared using 10 field and laboratory plume rise data bases (Peterson 1987). The numerical approach was superior in predicting the observed plume trajectories.

3.2 PLUME RISE IN NEUTRAL AND UNSTABLE CONDITIONS

Given the simplifications listed previously, the simplest form of the analytical solution for plume rise, Δh , can be obtained for neutral/unstable conditions where:

$$\Delta h = \left(\frac{3}{B^2} \frac{F_m x}{U^2} + \frac{3}{2B^2} \frac{F_b x^2}{U^3} \right)^{1/3}$$

where F_m and F_b are the *momentum and buoyancy fluxes* respectively and are defined as

$$F_b = R^2 \bar{w} g(T_g - T_a) / T_g$$

$$F_m = R^2 \bar{w}^2 T_a / T_g$$

The symbols are defined in Figures 3.1 and 3.2. Equation shows that Δh is a function of $(1/U)$; thus as wind speed increases, Δh decreases. In addition, Δh is proportional to $1/\beta$; thus as entrainment increases, Δh decreases (the greater the entrainment, the quicker the plume loses its initial buoyancy and momentum which reduces the plume rise). Finally, at larger downwind distances, the buoyancy term (second part of the equation) dominates the plume rise. In other words the contribution of the initial momentum to plume rise is quickly lost with increasing downwind distance.

Since the buoyancy term of the equation quickly dominates with increasing x , the first term is usually neglected for strongly buoyant sources. Plume rise is then given by:

$$\Delta h = \left(\frac{3F_b}{2\beta^2} \right)^{1/3} \frac{x^{2/3}}{U}$$

The trajectory of a buoyancy-dominated plume in a crosswind is a function of $x^{2/3}$. This behavior has been documented in numerous field studies. In addition, the entrainment constant, β , is consistently observed to be around 0.6 (see Briggs 1975, Table 3, for a list of observational studies). Observational studies in Alberta which support the $2/3$ power law behavior are listed in Table 3.1.

Djurfors and Netterville (1978) solved the conservation equations governing buoyant plume rise for the case of non-uniform wind conditions. A power law was used to represent the wind speed profile. The closed-form analytical solutions for both uniformly stable and neutral atmospheric conditions showed that the wind profile exponent has an equally strong effect on plume rise as the entrainment parameter, β . Verification of this method with observations awaits a data base with proper wind profile data.

If $F_b \ll F_m$, plume rise is a function of $x^{1/3}$ where

$$\Delta h = \left(\frac{3F_m}{\alpha^2 U^2} \right)^{1/3} x^{1/3}$$

Table 3.1 **Observational studies in Alberta of buoyant plume rise in neutral/unstable conditions.**

Location of Study	Findings and Comments	Reference
Gulf, Strachan Gulf, Rimbey Gulf, Nevis gas plants	2/3 power law behavior under neutral conditions	Lee et al. (1973) Whaley and Lee (1982)
Canoxy, Balzac gas plant	2/3 power law behavior under neutral conditions comparisons with Bosanquet et al. (1950), Lucas et al. (1963), Briggs (1969) plume rise equations	Kumar (1973)
Amoco, Vulcan gas plant	2/3 power law behavior under neutral conditions very buoyant flare stack source, radiant heat loss found to be 55% of available heat entrainment constant of 0.50 found	Leahey and Davies (1984)
Amoco, Crossfield gas plant	2/3 power law behavior under neutral conditions entrainment constant of 0.6 found	Davies et al. (1985)
Suncor (GCOS) tar sands plant	2/3 power law behavior under neutral conditions various analytical and numerical plume rise equations with different entrainment hypothesis were compared to large data base simplest analytical form had entrainment constant of 0.65	Slawson et al. (1978, 1980)

Table 3.1 **Observational studies in Alberta of buoyant plume rise in neutral/unstable conditions. (cont'd)**

Location of Study	Findings and Comments	Reference
wind tunnel modelling of Syncrude main stack	buoyant plumes follow the 2/3 power law	Neiman (1979) Winkel (1979)
	entrainment constant of 0.87 stack downwash, wind shear effects also investigated	Wilson (1979)
Shell, Waterton Gulf, Pincher Creek gas plants	final rise not observed	Whaley et al. (1980)
	2/3 power law behavior under near-neutral conditions	

This is the plume rise law for bent-over jets. Observational support for the $x^{1/3}$ behavior has been shown by a number of researchers (Keffer and Baines 1963, Patrick 1967, Sakiyama 1981). Since the entrainment rates for a bent-over jet are different from rates associated with a buoyant plume, empirical evidence yields $\alpha = 0.4 + 1.2/A$ where $A = \bar{w}/U$ (Briggs 1984). Examples of bent-over jets are ruptures from gas pipelines/wells and flares. In the case of gas pipeline/well ruptures, full-scale tests on ruptures from high-pressure gas lines were carried out near Waterton Park, Alberta and reported by Dielwart et al. (1979). Wilson (1979) and (1981) used the data to develop a model which predicts the behavior of the gas during initial release, expansion and plume rise. There were a number of unique factors that had to be accounted for such as density/compressibility effects and entrainment at the jet head. Final rise calculations based on a bent-over jet approach yielded remarkably good comparisons with $\alpha = 0.78$.

In the case of a flare, observation of the Acheson flare stack in Alberta provided an opportunity to compare a flame height and angle model developed by Leahey and Schroeder (1987). The flare stack flame behavior was assumed to be dominated by momentum effects where the rise follows an $x^{1/3}$ behavior. The model includes the effects of stack and gas parameters and meteorological conditions. Model predictions of flame height and angle resulted in linear correlation coefficients of 0.92 and 0.71 for flame height and angle respectively.

The spreading rate of a momentum or buoyancy-dominated plume in the initial rise phase is given by

$$R_p = \alpha, \beta \Delta h$$

Where entrainment constants (α, β) correspond to a jet and bent-over, buoyancy-dominated plume.

3.3 PLUME RISE IN STABLE CONDITIONS

For stable conditions, the solution of the conservation equations for plume trajectory yields a damped oscillatory behavior with a period equal to the Brunt-Vaisala period of $2\pi/N^2$ where N^2 is defined as:

$$N^2 = \frac{g}{T} \frac{\partial \theta}{\partial z}$$

$\frac{g}{T}$ acceleration due to gravity
 ambient temperature
 $\frac{\partial \theta}{\partial z}$ potential temperature gradient

Now Δh in stable conditions is given by:

$$\Delta h = \left(\frac{3}{U \beta^2 N^2} \right)^{1/3} \left(F_b \left(1 - \cos \frac{N x}{U} \right) + N F_m \sin N x / U \right)^{1/3}$$

and for a buoyancy-dominated plume

$$\Delta h = \left(\frac{3}{U \beta^2 N^2} \right)^{1/3} \left(F_b \left(1 - \cos \frac{N x}{U} \right) \right)^{1/3}$$

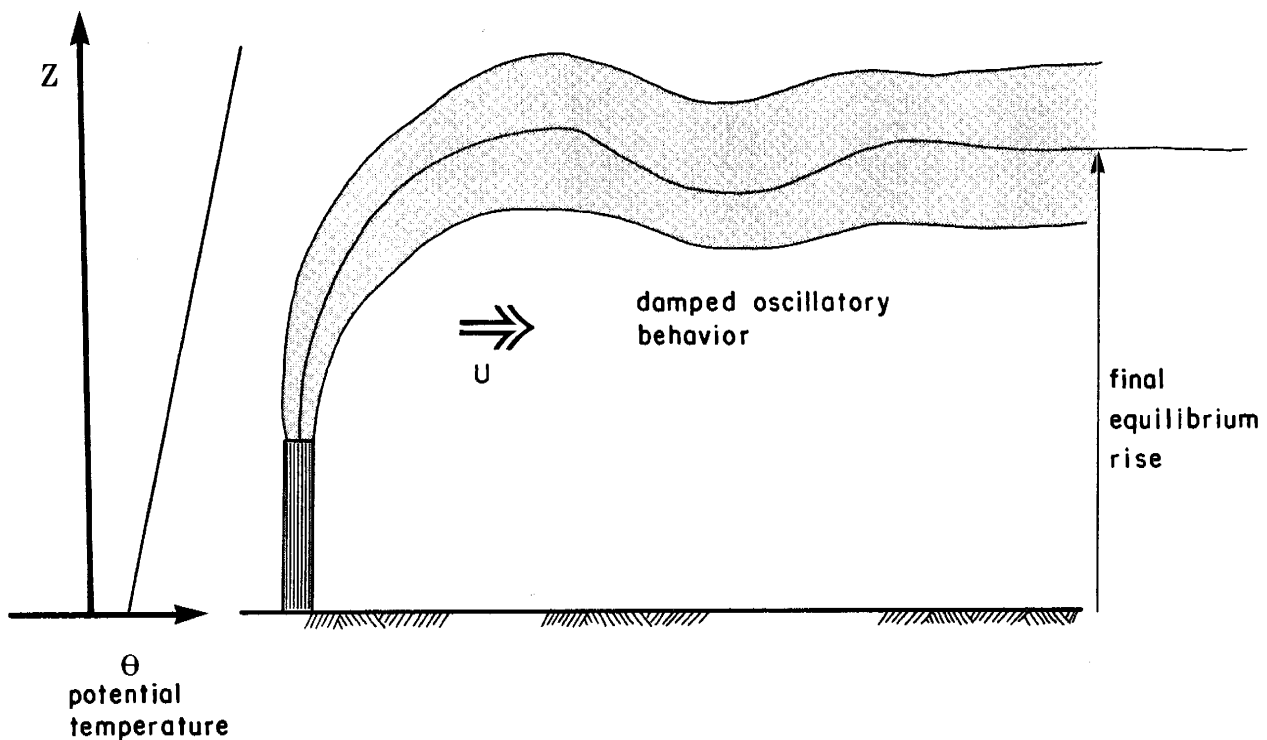


Figure 3.3 Depiction of plume rise in a stable atmosphere.

Table 3.2 Observational studies in Alberta of buoyant plume rise in stable conditions.

Location of Study	Findings and Comments	Reference
Suncor (GCOS) tar sands plant	plume oscillatory behavior clearly shown final rise comparisons to Briggs (1969) final rise equation showed poor agreement	Whaley and Lee (1978)
Suncor (GCOS) tar sands plant	various analytical and numerical plume rise equations with different entrainment hypotheses were compared to a large mixed stability data base a numerical model (N6) gave good trajectory and final rise predictions	Slawson et al. (1980)
Gulf, Strachan gas plant	one observational period under stable conditions no clear oscillatory behavior observed Briggs (1969) final rise equation consistently underpredicted observed final rise	Lee et al. (1973)

The trajectory of a bent-over plume in stable conditions is shown in Figure 3.3. This behavior has been observed in real situations, where plumes do reach a maximum rise and then find an equilibrium height (final rise). However, because of the drag effect of the atmosphere on the waves, no more than one or two oscillations are visible. Observations of plume rise in stable conditions in Alberta are not as common as those obtained under neutral/unstable conditions. Table 3.2 provides a list of these studies and their findings.

3.4 FINAL RISE IN NEUTRAL AND UNSTABLE CONDITIONS

The trajectory equation which describes the rise of a bent-over plume in neutral and unstable conditions yields $\Delta h \rightarrow \infty$ when $x \rightarrow \infty$. This obviously cannot be correct. The Equation only claims to describe plume behavior in the initial phase of plume rise; that is, where the plume self-generated turbulence dominates the mixing process. As the plume continues to grow and dilute, it gradually begins to "forget" its initial momentum and buoyancy. Ambient turbulence begins to dominate the mixing process and the plume "breaks up" and reaches final rise.

The determination of final rise in neutral/unstable conditions is the greatest gap in our knowledge of plume rise. The major problem is due to the lack of observational evidence to verify theories since, in neutral conditions, the visible plume disappears before reaching final rise and, in unstable conditions, the looping plume yields ambiguous determinations of final rise.

Numerous approaches to final rise appear in the literature. The simplest and most commonly used formula is the plume rise cut-off equation for buoyancy-dominated plumes (Briggs 1971). The trajectory equation is cut off at a final rise distance $3.5 x_*$ where:

$$x_* = 14 F_b^{5/8} \quad \text{for } F_b \leq 55 \text{ m}^4/\text{s}^3$$

$$x_* = 34 F_b^{2/5} \quad \text{for } F_b > 55 \text{ m}^4/\text{s}^3$$

There is no physical basis for the above equation; it is a purely empirical approach to this problem based on a number of observations of various power plant plumes.

This simple plume cut-off formula was later revised (Briggs 1975, 1984) to a more physically based formula. Since final rise under neutral and unstable conditions occurs when ambient turbulence dominates the dispersion of the plume, turbulent energy considerations were incorporated. Plume breakup occurs when the ambient turbulent dissipation rate (a measure of turbulent energy), equals or exceeds the plume dissipation rate. Mechanical and thermal turbulent energy are incorporated in the final rise equations. Table 3.3 lists the final rise equations using this approach.

A touchdown model can be used in convection-dominated atmospheres where up and down drafts occur (Briggs 1975, 1984). The concept of final rise is modified in this approach, since large downward eddies bring a plume to the ground before the plume rise terminates. The ambient downdraft velocity is characterized by, w_d , an average negative velocity. Table 3.3 lists the equations for the touchdown model.

Venkatram (1980) evaluated the effects of convection on plume behavior in the AOSERP study area. A model for application in convectively dominated boundary layers included the touchdown approach. High ground-level concentrations occurred under these conditions. Venkatram abandoned the idea of final rise under these conditions, and focused on predicting the impingement distance since the effort was directed on developing a model to predict maximum concentrations.

Benjamin (1975) incorporated the Briggs (1969) plume rise formula into a plume trajectory formula. The formula considers the effects of up and down drafts which characterize a convectively dominated mixed layer. The approach represents a departure from the conventional approach where concentrations are derived for unstable conditions by assuming a larger β . Benjamin found that the predicted maximum GLC's increased over those calculated for a neutral conditions. In addition, the effect of convection increases as the source buoyancy flux decreases.

Equations which describe turbulence limited final rise can be generated via the entrainment velocity route (Slawson and Csanady 1967, 1971). The assumption that entrainment is directly proportional to vertical velocity of the plume in the initial rise phase is replaced by different assumptions when ambient turbulence begins to affect the rise and spread of the plume. Central to this approach is the size of the plume (width)

Table 3.3 Final rise formulations.

Method	Neutral Conditions	Unstable Conditions
Turbulent energy approach	buoyant plume: $\Delta h = 1.2 \frac{F_b^{3/5}}{U u_*^2} (h_s + \Delta h)^{2/5}$	buoyant plume: $\Delta h = 4.3 (F_b/U)^{3/5} H^{-2/5}$
	momentum jet: $\Delta h = 0.93 \frac{F_m}{\alpha^2 U u_*^2} (h_s + \Delta h)^{1/7}$	momentum jet: $\Delta h = 2.3 (F_m/u)^{3/7} H^{-1/7}$
Touchdown model (unstable conditions only)		$\Delta h_d = 1.6 h_s \left(\frac{F_b}{U w_d^2 h_s} \right)^{1/3} (1 + 0.5 \Delta h_d / h_s)^{2/3}$ $\Delta h_d = h_s \left(\frac{3F_m}{\alpha^2 U w_d h_s^2} \right)^{1/3} [1 + (1 - \alpha/2) \Delta h_d / h_s]^{1/3}$ where: Δh_d = plume centreline above ground when lowest part of the plume impinges on ground

compared to the size of the energy-containing eddies. This was one of the early attempts to draw turbulence considerations into the problem of final rise. Wilson (1979) showed that this approach was in reasonable agreement with the final rise of the Syncrude plume as modelled in a wind tunnel.

Djurfors (1977b) incorporated the effects of atmospheric turbulence on plume behavior in an analytical expression. The advantage with his approach is that final rise can be determined without resorting to distinct final rise stages. The analytical expressions include the addition of a k factor, which is a function of atmospheric stability, ambient turbulence and the plume size. The final rise, which holds for all stabilities, is given by:

$$\Delta h_f = \left(\frac{3 (F_b + kF_m)}{U B^2 (N^2 + k^2)} \right)^{1/3}$$

Netterville (1983, 1990) continued this work and examined plume growth and breakup from an extrainment point of view. Atmospheric eddies are conceptualized to enter the plume, entrain plume mass, and then leave the plume with plume momentum and heat, thus "extrainment" occurs. Continued plume extrainment results in eventual plume breakup and final rise. The equations follow the same derivations as Djurfors, where the final rise equation is in the same form as the above equation, the only difference being the constant (i.e., instead of $3^{1/3}$, the value is 2.64). The analytical equations were compared to plume trajectory data taken in at the University of Alberta wind tunnel and are reported in Neiman (1979) and Winkel (1979). The comparisons were somewhat limited, but the new approach yields better estimates of plume behavior than the commonly accepted equations. Netterville (1983) acknowledges that more data are required to confirm these results. In addition, the value of k needs to be determined from turbulence measurements, since it is based on the turbulent intensity and length scale. Using plume rise observations from the Nanticoke, Ontario power plant, Netterville (1990) reported that the new approach significantly outperformed the traditional plume rise model at distances beyond 1 km from the source. The improvement in prediction comes at the price of a higher level of detail of inputs (temperature profile, intensity and scale of turbulence). However, this additional input can be obtained by SODAR and RASS remote sensors.

Final rise of buoyant plumes in unstable, low wind conditions can also be limited by the presence of an elevated inversion. For a plume that is fully trapped (i.e., unable to penetrate the capping inversion), final rise is given by (Weil 1979):

$$\Delta h = 0.62 h'$$

where h' is the distance between stack top and the capping inversion.

Observational studies of final rise of plumes in Alberta during neutral/unstable conditions are given in Table 3.4.

3.5 FINAL RISE IN STABLE CONDITIONS

During stable conditions, final rise occurs as a result of the negative buoyancy acquired by entraining ambient air and carrying it to levels of higher ambient potential temperature. For a buoyant plume the final rise is taken to be the equilibrium height of the oscillatory behavior described by the equation given earlier. The final rise of a buoyant plume in stable conditions is given by:

$$\Delta h = 2.6 \left(F_b / UN^2 \right)^{1/3}$$

A similar formula for final rise of bent-over jets is not available due to a lack of observational data. The constant, 2.6, is based on experimental evidence. Observations of final rise of plumes in Alberta during stable conditions are summarized in Table 3.2.

3.6 PLUME PENETRATION INTO ELEVATED INVERSIONS

Rising plumes often encounter a capping inversion, especially plumes from hot industrial sources in neutral and unstable atmospheres. If a plume can totally penetrate into this layer, the layer will prevent downward mixing and ground-level concentrations will be quite low. On the other hand, if the plume is unable to penetrate the inversion, then the adverse condition of trapping occurs where all the plume material is trapped within the mixed layer, dramatically increasing the ground-level concentrations. Between these two extremes, the plume may only partially penetrate the inversion. These situations are illustrated in Figure 3.4.

Table 3.4 Observations in Alberta of final rise in neutral/unstable conditions.

Location	Findings and Comments	Reference
Suncor (GCOS) tar sands complex	agreement with the final rise limiting value of Briggs (1969)	Whaley and Lee (1978)
Suncor (GCOS) tar sands complex	comparisons made with various rise formulations Briggs (1969, 1971, 1972), Montgomery et al. (1971, 1972), Holland (1953), Brummage (1968), Moses and Carson (1967) no clear winner was found layered stability (elevated inversions were common)	Fanaki (1978), Fanaki et al. (1979)
Suncor (GCOS) tar sands complex	final rise observed was consistently higher than the Briggs (1969) limiting value) semi-empirical final rise formula developed for neutral conditions	Slawson et al. (1978, 1980)
Waterton pipe burst experiments	momentum dominated release final rise agreement with Briggs (1975) turbulent dissipation approach for jet final rise	Dielwart et al. (1979) Wilson (1981)
Wind tunnel modelling of Syncrude main stack	final rise compared with Briggs (1975) turbulent dissipation approach for buoyant plumes and the Slawson and Csanady (1971) entrainment approach the Briggs (1975) approach overpredicted final rise	Neiman (1979) Winkel (1979) Wilson (1979)

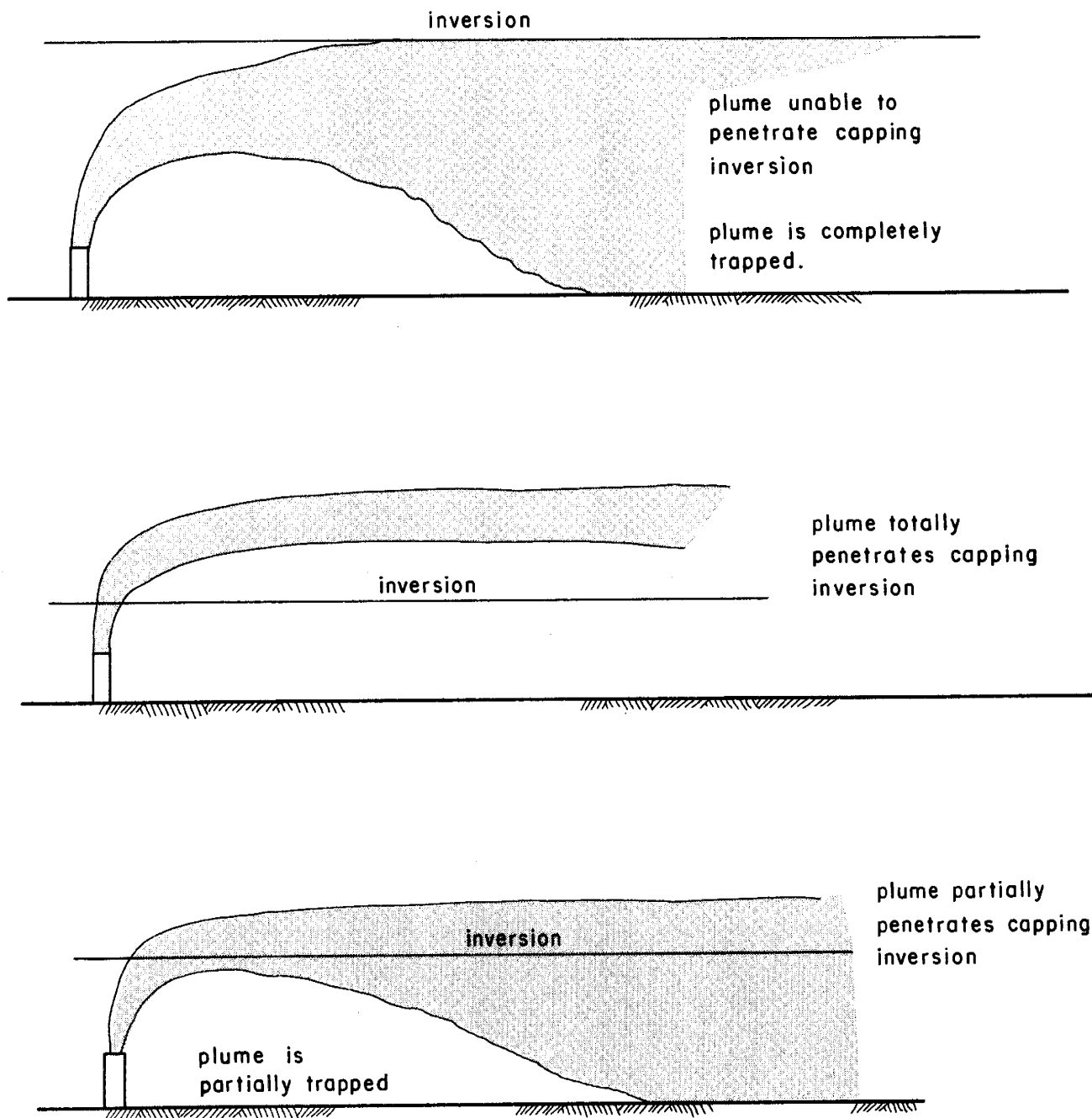


Figure 3.4 Plume penetration of elevated inversions.

There are numerous approaches to predict the degree of plume penetration given in the literature; unfortunately there is a scarcity of observational data to evaluate the different approaches although water flume modelling has helped considerably (Manins 1979). In addition, depending on the characteristic of the profile, i.e., deep vs. shallow capping inversion, different equations apply (see Briggs 1984 for review and recommendations).

A simple formula to predict plume penetration was proposed in Kumar and Djurfors (1978) and later detailed and tested against observational data in Kumar (1980). Forty nine, time-mean plumes of the Suncor plume were used as an evaluation data base from which Kumar concluded that the fraction of plume depth entering the inversion is given by:

$$P = \frac{1+\alpha}{2\alpha} - \frac{h}{2\alpha Z}$$

α entrainment parameter (0.6)
 h mixed layer height above stack top
 Z plume rise at the point of penetration.

3.7 STACK AERODYNAMIC DOWNWASH

Stack aerodynamic downwash is a special situation where stack emissions are drawn into the low pressure area on the lee side of the stack. The occurrence of this phenomenon depends on a number of factors including the shape of the stack tip, the exit momentum of the plume and the wind speed. If downwash occurs, high ground-level concentrations can occur (Briggs in Noll and Davis, 1976) since the plume is effectively much closer to the ground. This behaviour is shown in Figure 4.5.

Early wind tunnel work (Sherlock and Stalker 1941) lead to the famous 1.5 rule of thumb (Briggs 1969): downwash will occur if the exit velocity, $\bar{w}$, is less than 1.5 times the stack top windspeed, U . The universally accepted formula to correct for decreased plume rise under downwash conditions is given by (Briggs in Noll and Davis 1976):

$$H_{\text{down}} = 2 (\bar{w}/U - 1.5) D$$

where D is the internal stack diameter. This equation is added algebraically to the stack height, h_s , to produce a virtual stack height, h_v . The effective stack height is given by

$$h_{\text{eff}} = h_v + \Delta h$$

These equations are commonly used in dispersion models to account for this effect. Further observation of this phenomenon (wind tunnel - Wilson 1979; field observations - Rittman 1982) and a re-examination of the Sherlock and Stalker data suggested that the use of these equations is questionable (Halitsky 1978). With respect to the onset of downwash, different approaches have been suggested. Based on wind tunnel observations of the Syncrude stack and from momentum flux considerations, Wilson (1979) suggested that downwash occurs when $\phi_M < 2$ where:

$$\phi_M = (\rho_s/\rho_a)(\bar{w}/U_s)^2 C$$

where

ρ_s/ρ_a are the density of the stack gas and atmosphere respectively,
 U_s is the wind speed at stack height
 C is a correction factor required to match momentum rise in the wind tunnel to full scale.

An alternate criterion based on nine wind tunnel and towing tank observations of buoyant sources (Overcamp 1983) requires the definition of a modified speed ratio, MSR, (which is the same as the square root of Wilson's ϕ_M with $C = 1$) and the stack Froude number where:

$$Fr_s = \bar{w} / [g R (\rho_a - \rho_s) / \rho_s]^{1/2}$$

The downwash/non-downwash regions were defined by the line on the graph shown on Figure 3.6.

Alternate downwash correction factors have been proposed. Wilson (1979) suggested the use of his ϕ_M where

Δh (with downwash)	$= \Delta h$	for $\phi_M > 2.0$
	$= \Delta h (1 - \phi_M)$	for $1.0 < \phi_M < 2.0$
	$= 0$	for $\phi_M < 1.0$

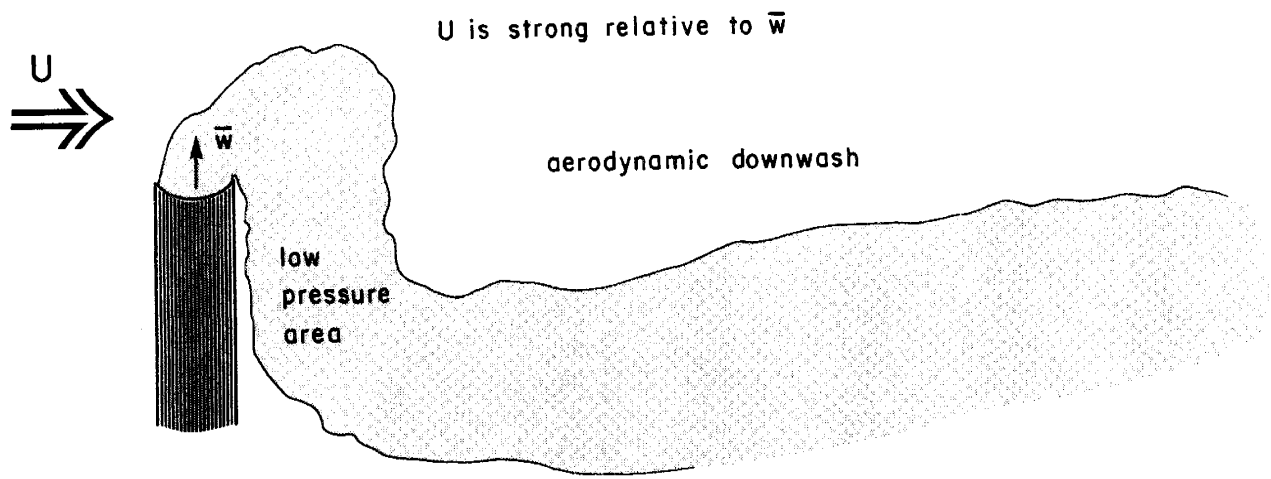


Figure 3.5 Depiction of stack aerodynamic downwash.

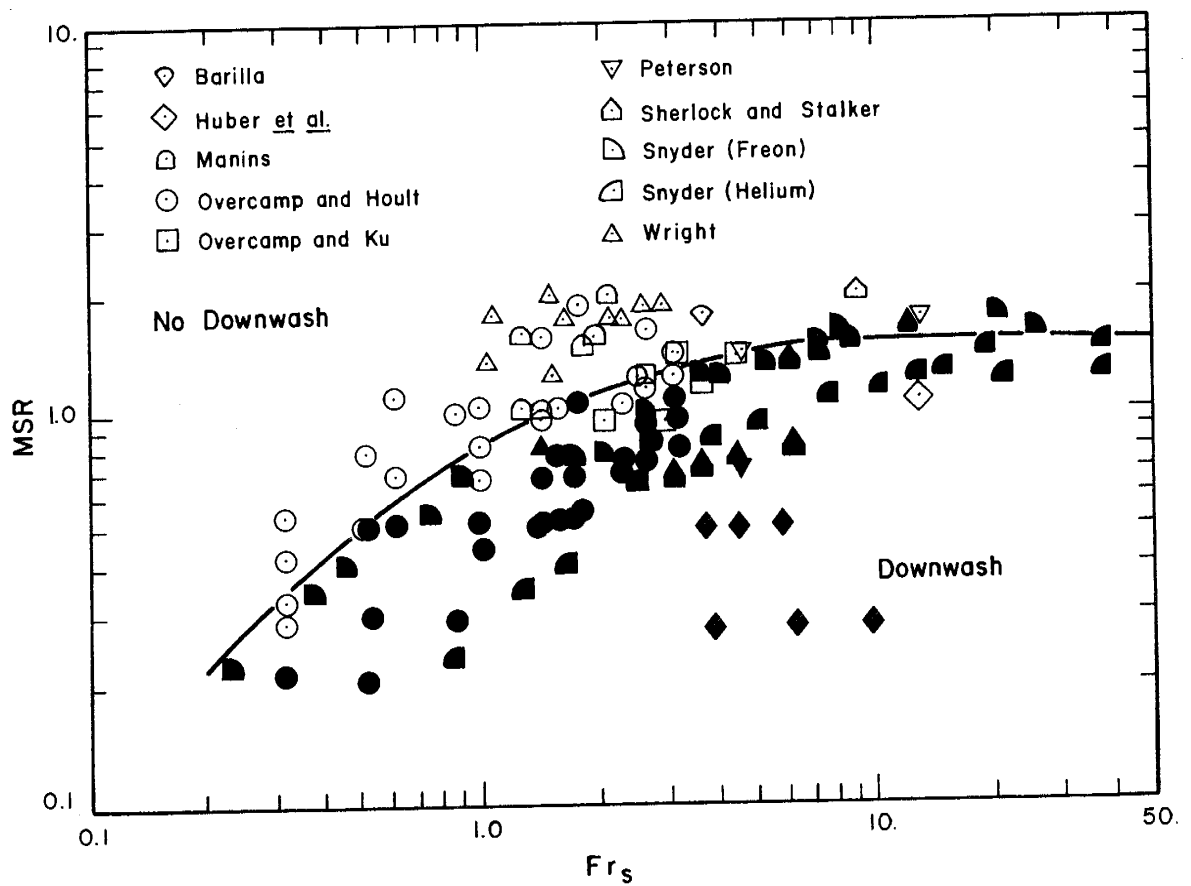


Figure 3.6 Modified speed ratio, MSR, vs. source Froude number, Fr_s . Plot shows zones of downwash/non-downwash occurrences (Overcamp 1983).

On the basis of the growth of the turbulent wake in the lee of the stack, another correction method (Briggs 1984) is given by:

$$\Delta h \text{ (with downwash)} = \Delta h - A (2R_x)^{1/2}$$

where A is a constant of proportionality. Davies et al. (1985) conducted a field study during the winter of 1984/85 to obtain plume observations of the buoyancy-dominated plume from the Crossfield gas plant (north of Calgary). Under downwash conditions, there was good agreement between observations and predictions.

3.8 THE BEHAVIOUR OF MOIST PLUMES

Plumes that contain a large amount of water (such as cooling tower plumes and scrubber plumes) have a potential to lose or gain substantial buoyancy due to the release or uptake of latent heat of evaporation, a thermodynamic aspect that several investigators have examined (Morton 1957, Csanady 1971, Wigley and Slawson 1971, Weil 1974, Schatzmann and Policastro 1984).

If the plume contains liquid water in the form of small droplets, and if the plume rises and mixes with dry air, these droplets can evaporate. The phase change uses latent heat from the air and plume and thus the plume cools and loses buoyancy. In effect, the plume could impinge on the ground sooner than a dry plume in equivalent atmospheric conditions.

If the plume contains water vapor and rises into moist or very cold air, some of the water vapor may condense thus releasing latent heat. This will warm the plume and thus increase plume buoyancy. However, it has been shown (Wigley and Slawson 1971, Briggs 1984) that even under the most favorable atmospheric conditions, the increase in plume rise due to condensation is very small.

Djurfors (1977c) investigated the situation of wet scrubbing on an SO₂ plume. Wet scrubbing is a process that removes SO₂ from the effluent gases and would appear to be beneficial in decreasing downwind SO₂ concentrations. However, wet scrubbing also reduces the effluent temperature considerably (and thus reduces the effective release

height) which causes an increase in ground-level concentrations. On the other hand, the addition of large amounts of water to the effluent can result, under certain conditions, in extra plume rise due to a release of latent heat. Considering all these effects Djurfors concluded that no substantial gain in air quality could be made by wet scrubbing unless removal efficiencies are maintained at greater than 90%. The effect of condensation on increasing plume rise was seen to be very small.

3.9 PLUME RISE FROM GAS TURBINES AND RECIPROCATING INTERNAL COMBUSTION ENGINES

The great majority of observational work on plume rise has focussed on rise from large, very buoyant industrial sources. This data base has been used in the development and verification of the plume rise models described earlier. Another problem of relevance to Alberta, is the plume rise from compressor station stacks. These sources, unlike the very buoyant, large industrial sources, are characterized by short stacks (about 5-10 m), high exit temperatures (500-900 K) and high exit speeds (20-80 m/s). They are also located near buildings of about the same height as the stack. Thus it is a concern as to whether plume rise formulations used for industrial sources apply to the compressor-station-type stacks.

Some work has been done (Briggs 1984) to address this question. He reviewed three wind tunnel/field measurement experiments (Moore and Robins 1975, Hoydysh et al. 1975, and Hoult et al. 1975) and concluded that the rise from these types of sources is consistent with the predictions of the standard plume rise equations for buoyant plumes. However, building downwash can reduce the plume rise depending on the ratio of exit velocity to wind speed.

Observations of plume rise from a gas turbine and three reciprocating engines were compared to the results given by existing plume rise models (Balentine et al. 1983). The Briggs buoyant plume rise equation underpredicted the observed rise; however, the Briggs combined momentum and buoyant plume rise equation tended to slightly overpredict the turbine plume rise. A modified five-parameter model was developed which gave good agreement to all of the data. This model has the same theoretical basis; however, five constants were included, the values of which were determined by a regression on the data. The effect of building downwash on plume rise was not explicitly included.

3.10 SUMMARY

Plume rise models are based on the conservation equations and a closure assumption. The solutions have led to simple power law expressions to describe the trajectory of both buoyancy- and momentum-dominated plumes in different atmospheric conditions. Close to the source, plume rise increases as the two-thirds power of the distance. The $2/3$ power law has been confirmed in Alberta under near-neutral atmospheric conditions for a variety of sources. Entrainment constants ranging from 0.5 to 0.9 have been observed.

Final rise has been studied theoretically, but few data are available for comparison because of the difficulty of making observations at larger distances from the source. Alberta studies of final rise for stable conditions generally found poor agreement with the Briggs (1969) equation, while observations of final rise for neutral/unstable conditions gave mixed results.

Similarly, plume penetration of elevated inversions has been studied theoretically, but observations in Alberta are sparse. With elevated layers present 33% of the time and shallow mixing layers in the colder months, this phenomenon is expected to be fairly common. Ground-level concentrations will be reduced in proportion to the fraction of plume entering the inversion layer.

Low stack exit speeds increase the potential for aerodynamic downwash. The simple, commonplace rules of thumb are inadequate to describe the onset of downwash or its effect on ground-level concentration. More advanced plume trajectory correction methods are supported by the limited field observations in Alberta.

3.11 REFERENCES

- Balentine, H.W., H.J. Williamson, S.M. Zubrick, S.P. Coerr. 1983 Plume rise from gas turbines and reciprocating internal combustion engines in gas transmission. Radian Corporation DCN 83-240-251-11, Project PR-156-139. Austin, Texas, 204 pp.
- Batchelor, G.K. 1954. Heat convection and buoyancy effects in fluids. *Quart. J. Roy. Meteor. Soc.* 80(345): 339-358.
- Benjamin, S.F. 1975. The behavior of hot plumes from tall chimneys under conditions of strong thermal convection. Presented at PNWIS-APCA Meeting, Nov. 1975, Vancouver, British Columbia. 41 pp.
- Bosanquet, C.H. 1957. The rise of a hot waste gas plume. *J. Int. Fuel* 30: 322 - 328.
- Bosanquet, C.H., W. F. Carey and E. M. Halton. 1950. Dust deposition from chimney stacks. *Proc. Inst. Mech. Eng.* 162: 355 - 367.
- Briggs, G.A. 1984. Plume rise and buoyancy effects: in *Atmospheric science and power production*, Chapt. 8: 327 - 366, D. Randerson (Editor). U.S. Dept. of Energy, DOE/TIC-27601. (DE84005177)
- Briggs, G.A. 1975. Plume rise predictions: in *lectures on air pollution and environmental impact analyses*. Chap. 3: 59 - 111. Boston, Sept. 29 - Oct. 3, American Meteorological Society, Boston, Mass.
- Briggs, G.A. 1972. Discussion. Chimney plumes in neutral and stable surroundings. *Atmos. Environ.* 6: 507 - 510.
- Briggs, G.A. 1971. Some recent analyses of plume rise observations. Pages 1029 - 1032 in *Proceedings, Second International Clean Air Congress*. Dec. 6-11, 1970. Washington. Academic Press.
- Briggs, G.A. 1969. Plume rise. US Atomic Energy Commission Critical Review Series. Available as TID 25075 from NTIS. Springfield, VA. 81 pp.
- Brummage, K.G. 1968. The calculation of atmospheric dispersion from a stack. *Atmos. Environ.* 2: 197 - 224.
- Csanady, G.T. 1971. Bent-over vapor plumes. *J. Appl. Meteor.* 10: 36 - 42.
- Davies, M.J.E. 1986. Plume visualization techniques. Report prep. for Standards and Approvals Division, Alberta Environment. 27 pp.
- Davies, M.J.E., D. D. Kristmanson and R. D. Rowe. 1985. A field investigation of stack aerodynamic downwash. Prepared for Standards and Approvals Division of Alberta Environment, Edmonton, Alberta. 92 pp.
- Davidson, W.F. 1949. The dispersion and spreading of gases and dust from chimneys. *Trans. Conf. on Industrial Wastes, 14th Ann. Meeting, Ind. Hygiene Found.* 38 - 55 (Nov. 18, 1949).

- Dielwart, J., H. Knox and R.P. Angle. 1979. Hydrogen sulphide isopleth prediction - phase II: pipe burst study. Alberta Petroleum Industry Government Environmental Committee (APIGEC). 45 pp. plus appendices.
- Djurfors, S. 1977(a). On the rise of vertical plumes. Syncrude Canada Ltd. Professional Paper. 1977-8. 14 pp.
- Djurfors, S. 1977(b). On the rise of buoyant plumes in turbulent environments. Syncrude Canada Ltd. Professional Paper, 1977-4. 10 pp.
- Djurfors, S. 1977(c). Effect on plume behavior of cooling by wet scrubbing. Engineering Digest May. 35 - 39.
- Djurfors, S. and D. Netterville. 1978. Buoyant plume rise in non-uniform wind conditions. J. Air Poll. Cont. Assoc. 28: 780 - 784.
- Fanaki, F. Compiler. 1978. Meteorology and air quality winter field study in the AOSERP study area, March 1976. Prep. for the Alberta Oil Sands Environmental Research Program by Atmospheric Environment Service. AOSERP Report 27. 249 pp.
- Fanaki, F., R. Hoff and M. Lusi. 1979. Air pollution studies at the Alberta oil sands area. PNWIS-APCA meeting, Edmonton, Alberta. Paper 79 - 94.
- Halitsky, J. 1978. Comment on a stack downwash prediction formula. Atmos. Environ. 12: 1575 - 1576.
- Hewett, T.A., J.A. Fay and D.P. Hoult. 1971. Laboratory experiments of smokestack plumes in a stable atmosphere. Atmos. Environ. 5: 767 - 789.
- Holland, J.Z. 1953. A meteorological survey of the Oak Ridge area. U.S. Atomic Energy Commission Report ORO-99, 554 - 559.
- Hoult, D.P., S.R. O'Dea, G.L. Touchton and R.J. Ketterer. 1975. Turbulent plume in a turbulent crossflow: comparison of wind tunnel tests with field observations. 68th Annual Meeting of APCA. (preprint No. 75-49.1).
- Hoydysh, W.G., P. Doisey and T. Yoshida. 1975. Scaling considerations for simulating plume rise and plume dispersion from gas turbine generators. 68 Annual Meeting of APCA. (preprint No. 75-49.5). June 15-20, Boston.
- Keffer, J.F. and W.D. Baines. 1963. The round turbulent jet in a crosswind. J. Fluid Mech. 15: 481-496.
- Kumar, A. 1980. Some observations of plume penetration in the Athabasca Tar Sands area. Syncrude Canada Limited Report. 13 pp.
- Kumar, A. and S.G. Djurfors. 1978. A model to predict violation of clean air regulations. Proceedings 4th Joint Conference on Sensing of Environmental Pollutants: 281 - 288 (Syncrude Canada Ltd. Professional Paper 1977-5).
- Kumar, V K. 1973. Sulphur recovery plant plume rise. M.Sc., Thesis. University of Calgary. Dept. of Mechanical Engineering. 103 pp.

- Leahey, D.M. and M.B. Schroeder. 1987. Observations and predictions of jet diffusion flame behavior. *Atmos. Environ.* 21(4): 777 - 784.
- Leahey, D.M. and M.J.E. Davies. 1984. Observation of plume rise from sour gas flares. *Atmos. Environ.* 18(5): 917 - 922.
- Lee, G.K., H. Whaley and J.G. Gainer. 1973. Plume dispersion research at natural-gas sulphur-extraction plants. Canadian Department of Energy, Mines and Resources - Mines Branch, Ottawa. Research Report R 265 35 pp.
- Lucas, D.H., D. J. Moore and G. Spurr. 1963. The rise of hot plumes from chimneys. *Int. J. Air Water Pollut.* 7: 473 - 500.
- Manins, P.C. 1979. Partial penetration of an elevated inversion layer by chimney plumes. *Atmos. Environ.* 13: 733 - 741.
- Montgomery, T.L., S.B. Carpenter, W.C. Colbaugh, and F.W. Thomas. 1972. Results of recent TVA investigations of plume rise. *J. Air Poll. Cont. Assoc.* 22: 779 - 784.
- Montgomery, T.L., S.B. Carpenter, W.C. Colbaugh, and F.W. Thomas. 1971. Fullscale study of plume rise at large electric generating stations - Bull Run supplement. Air Quality Branch, Tennessee Valley Authority. Muscle Shoals, Ala. 58 pp.
- Moore, D.J. 1965. On the rise of hot plumes in the atmosphere. *Int. J. Air Water Poll.* 9: 233 - 237.
- Moore, D.J. and A.G. Robins. 1975. A comparison between calculated and observed full scale and model gas turbine plume trajectories. 68th Annual Meeting of APCA. (preprint No. 75-49.4). June 15-20, Boston.
- Morton, B.R. 1957. Buoyant plumes in a moist atmosphere. *J. Fluid Mech.* 2: 127 - 144.
- Morton, B.R., Taylor, G.I. and J.S. Turner. 1956. Turbulent gravitational convection from maintained and instantaneous sources. *Proc. Royal Soc. A234*: 1 - 23.
- Moses, H. and J.E. Carson. 1967. Stack design parameters influencing plume rise. Paper presented at the 60th Annual Meeting of the Air Pollution Control Association. Cleveland, Ohio. Paper 67 - 84.
- Moses, H. and G.H. Strom. 1961. A comparison of observed plume rises with values obtained from well-known formulas. *Jour. Air Poll. Cont. Assoc.* 11(10): 455 - 466.
- Neiman, O. 1979. Effects of man-made obstacles on plume dispersion at a surface mining site. M. Sc. Thesis, Mech. Eng., University of Alberta. 127 pp.
- Netterville, D.D.J. 1990. Plume rise, entrainment and dispersion in turbulent winds. *Atmos. Environ.* 24A(5):1061-1081
- Netterville, D.D.J. 1983. Plume rise in turbulent winds. Syncrude Canada Ltd., Research and Development Report. 29 pp.

- Noll, K.E. and W.T. Davis, Editors. 1976. G. A. Briggs - Estimation of downwash effects Chapt. 6 in Power Generation: Air pollution monitoring and control. Ann Arbor Science, Ann Arbor, MI. 65 - 70.
- Overcamp, T.J. 1983. Guidelines for stack downwash in plume modeling experiments. Paper presented at 76th Annual Meeting of the Air Pollution Control Association, Atlanta, Georgia, June 19 - 24, 1983. Paper No. 36.3, 11 pp.
- Patrick, M.A. 1967. Experimental investigation of mixing and penetration of a round turbulent jet injected perpendicularly into a transverse stream. Trans. Inst. Chem. Eng. (London), 45: 16 - 31.
- Peterson, R. L. 1987. Performance evaluation of integral and analytical plume rise algorithms. Jour. Air Poll. Cont. Assoc. 37: 1314 - 1319.
- Rittman, B. E. 1982. Application of two-thirds law to plume rise from industrial-sized sources. Atmos. Environ. 16(11): 2575 - 2579.
- Sakiyama, S.K. 1981. Plume dispersion modelling at the Whiteshell Nuclear Research Establishment. MASc Thesis, Department of Mechanical Engineering, University of Waterloo. 219 pp.
- Scorer, R. S. 1959. The rise of bent-over hot plumes. Adv. in Geophysics, 6: 399 - 411, The Academic Press.
- Schatzmann, M. 1979. An integral model of plume rise. Atmos. Environ. 13: 721 - 731.
- Schatzmann, M. and A. J. Policastro. 1984. Plume rise from stacks with scrubbers: a state of the art review. Bull. Amer. Meteor. Soc. 65 (3): 210 - 215.
- Sherlock, R. H. and E. A. Stalker. 1941. A study of flow phenomena in the wake of smoke stacks. Eng. Research Bulletin No. 29, University of Michigan, Ann Arbor, MI.
- Slawson, P. R., G. A. Davidson and C. S. Maddukuri. 1980. Dispersion modelling of a plume in the tar sands area. Syncrude Environmental Research Report 1980-1. 316 pp.
- Slawson, P. R., G. A. Davidson, W. McCormick, G. Raithby. 1978. A study of the dispersion characteristics of the G.C.O.S. plume. prepared by Envirodyne Ltd. for Syncrude Canada Limited. 113 pp.
- Slawson, P. R. and G. T. Csanady. 1971. The effect of atmospheric conditions on plume rise. J. Fluid Mech, 47: 33 - 49.
- Slawson, P. R. and G. T. Csanady. 1967. On the mean path of buoyant bent-over chimney plumes. J. Fluid Mech. 28: 311 - 322.
- Snyder, W. H. 1974. Fluid modelling program of the meteorology laboratory, U.S. Environmental Protection Agency. 5th Meeting. NATO/CCMS Expert Panel on Air Pollution Modelling. Danish Atomic Energy Commission Research Establishment, Risø, Roskilde, Denmark, 4 - 6 June 1974.
- Sutton, O.G. 1947. The theoretical distribution of airborne pollution from factory chimneys. Quart. J. Roy. Meteor. Soc. 73: 426 - 436.

- Taylor, G.I. 1948. Dynamics of a mass of hot gas rising in air. USAEC Report MDDC-919 (LADS-276) Los Alamos Scientific Laboratory, NTIS.
- Venkatram, A. 1980. Evaluation of the effects of convection on plume behavior in the AOSERP study area. Prep. for the Alberta Oil Sands Environmental Research Program. AOSERP Report 95. 75 pp.
- Weil, J. C. 1979. Assessment of plume rise and dispersion models using lidar data. Report PPSP-MP-24, Maryland Power Plant Siting Program.
- Weil, J. C. 1974. The rise of moist, buoyant plumes. J. Appl. Meteor. 13: 435 - 443.
- Whaley, H. and G. K. Lee. 1982. The behavior of buoyant plumes in neutral and stable conditions in Canada. Atmos. Environ. 16(11): 2555 - 2564.
- Whaley, H. and G. K. Lee. 1978. An assessment of plume dispersion parameters measured in fall and winter at a Tar-Sands refinery complex. Jour. Air Poll. Cont. Assoc. 28(6): 589 - 593.
- Whaley, H., G. K. Lee and J. G. Gainer. 1980. The behavior of buoyant merging plumes in the Rocky Mountain foothills. Jour. Air Poll. Cont. Assoc. 30(7): 782 - 786.
- Wigley, T. M. L. and P. R. Slawson. 1971. On the condensation of buoyant, moist, bent-over plumes. J. Appl. Meteor. 10: 253 - 259.
- Wilson, D. J. 1981. Expansion and plume rise of gas jets from high pressure pipeline ruptures. Prepared for Alberta Environment, Pollution Control Division, Contract No. 810786. 61 pp.
- Wilson, D. J. 1979. Wind tunnel simulation of plume dispersion at Syncrude Mildred Lake site. Final report of Contract 98-0117 EA, Syncrude Environmental Affairs Dept. Syncrude Canada Ltd. 198 pp.
- Winkel, G. R. 1979. Wind tunnel modelling of plume rise and dispersion. M.Sc. Thesis, Mech. Eng., University of Alberta. 135 pp.

4.0 **ATMOSPHERIC DISPERSION OVER GENTLE TERRAIN**

The dispersion of airborne material by the natural wind is a very complex phenomenon. The complexity is in large part due to the wide spectrum of eddy sizes present in turbulent atmospheric flow. Even under the idealized conditions of a laboratory the dynamics of turbulence and turbulent diffusion are generally regarded as the most intractable problems in the whole of fluid mechanics at the present time. The problems are even more complicated in the atmosphere because of the pronounced effect of temperature stratification and the frequently non-uniform and irregular nature of the earth's surface. Therefore, it should not be surprising that no general complete theory has been formulated to express the physical relationship between ambient concentrations of air pollutions and the causative meteorological factors and processes.

A large body of empirical data on atmospheric dispersion has been accumulated from various activities. Practical and/or semi-theoretical expressions and working equations derived from field studies for limited classes of situations, have proved useful in the solution of air pollution problems. All currently available prediction schemes refer to rather idealized conditions of flow and terrain from which there are large deviations in practice.

The fundamental probabilistic nature of atmospheric behavior must also be recognized. There are random variations in the airflow and in the apparent dispersive ability of the atmosphere which arise because the flow is turbulent with a wide range of eddy sizes. This implies that over any given scale of interest dispersion estimates can, at best, provide information only on statistical quantities. On individual occasions the actual results may depart markedly from predicted values.

It is generally agreed that modelling for the practical estimation of dispersion can provide useful results that could be very difficult or expensive to obtain by other means. In unusually complex situations, it will always be desirable to check the estimates obtained by appropriate field observations.

The main characteristics of the three current working theories of diffusion are summarized in Table 4.1 (Pasquill 1971, 1974, 1975). The Gaussian model is unique in that it can be related to all three approaches. It is simple enough that one can visualize diffusion effects and also flexible enough to incorporate a host of special phenomena. For these reasons the rest of this chapter will talk in terms of the Gaussian model. It is implicit that the other approaches could also be used, at least within their limitations.

4.1 THE GAUSSIAN PLUME MODEL

Gaussian diffusion formulas first appeared as solutions to the gradient transfer equations (Roberts 1923, Sutton 1932, Frenkiel 1953). Later they were derived from statistical theory (Sutton 1953) and used as mathematical diffusion models to correlate empirical data (Cramer 1957; Hay and Pasquill 1957; Gifford 1960a, 1961).

The Gaussian function can provide a general description of average plume diffusion by virtue of the central limit theorem of statistics (Batchelor 1949) and the essentially random nature of the phenomenon. For small diffusion times the distribution of particles should take the same form as the wind fluctuation distribution since the particle trajectories coincide with the instantaneous wind (Taylor 1921, Lin and Reid 1963). Turbulent velocities in the atmosphere approximate a Gaussian distribution fairly closely. Also, if the Fourier components of the velocity time series are assumed to be independent random variables (principle of random phase), then the resulting distribution can be shown to be Gaussian (Csanady 1973). Strictly speaking, individual Fourier components in turbulent flow are not independent. However, much successful work has been done assuming that only the Fourier components of comparable wavelength are interrelated. Presumably there are enough independent components to produce the same result as if all such components were independent. Although "the justification for the Gaussian nature of the particle distribution remains vague," the foregoing non-rigorous arguments "at least suggest that major departures are unlikely" (Csanady 1973).

Table 4.1 Current working theories of diffusion (after Pasquill 1975).

Nature of theory	Statistical	Gradient-transfer (a) surface layer (b) convective	Similarity scaling
Basic quantities	RMS eddy velocity and Lagrangian autocorrelation	eddy diffusivity	(a) friction velocity, heat flux (b) heat flux, convective mixing depth
Limitations	homogeneity stationarity	small-scale action	(a) surface layer (b) unstable conditions
Data Required	spectrum of turbulence	profile of wind and diffusivity	profiles of wind and temperature (a) close to ground (b) through planetary boundary layer
Dispersion property	horizontal/vertical spread (elevated source) horizontal spread (ground source)	vertical spread (ground source)	(a) vertical spread (ground source at short range) (b) vertical and horizontal spread in unstable conditions

Table 4.1 Current working theories of diffusion (after Pasquill 1975). (cont'd)

Nature of theory	Statistical	Gradient-transfer (a) surface layer (b) convective	Similarity scaling
Advantages	uses actual turbulence based on simple kinematic and statistical relations	gives useful results even outside theoretical range of applicability allows for spatial and temporal changes provides opportunity to use high-speed computers for numerical solutions	embodies current knowledge of boundary layer
Disadvantages	fundamental quantities difficult to measure	difficult to specify K in objective manner utility of results not known beforehand (called semi-empirical method by Russian authors)	fundamental inputs difficult to measure relationships between interdependent variables may be suspect interpolations required to bridge transition between regimes

In practice, deviations from a Gaussian distribution are common, but in an apparently non-systematic way. The concentration profiles through a plume may be too peaked, too flat, or too skewed; but no single class of functions seems to adequately describe all possibilities. Thus it is practical to maintain the Gaussian interpolation formulas and inquire into the magnitude of possible resulting errors (Islitzer and Slade 1968). In the planetary boundary layer when the wind is strongly sheared and skewed, the Gaussian assumption is faulty in principle, but even in this case it is still a good approximation for short travel times (Gifford 1975). For elevated sources the concentration distribution tends to be Gaussian, but for ground-level sources, a different vertical distribution is indicated. However, departures from the Gaussian distribution are not of practical significance (Pasquill 1974).

In a major Alberta tracer study, the ERCB (1990) reported that a large portion of their 846 plume traverses were reasonably approximated by a Gaussian profile, many were virtually indistinguishable, and a few were skewed or bi-modal. Skewness coefficients were normally distributed about zero, indicating that departure from the Gaussian shape are random. Kurtosis coefficients showed that many observed profiles were slightly less peaked than the Gaussian; however, on average there was good agreement with the Gaussian shape.

The virtues of the Gaussian distribution function are considerable: statistically it is completely determined by its second moment; mathematically it possesses a self-reciprocal Fourier transform (Gifford 1968); moreover, it agrees reasonably well with the bulk of experimental data (Islitzer and Slade 1968, Pasquill 1974).

"The important point is that the Gaussian formula, properly used, is peerless as a practical diffusion modelling tool. It is mathematically simple and flexible, it is in accord with much though not all of working diffusion theory, and it provides a reliable framework for the correlation of field diffusion trials as well as the results of both mathematical and physical diffusion modeling studies." (Gifford 1975).

The basic working formula of the Gaussian plume diffusion model can be viewed in at least three ways: (a) it can be thought of as a formal solution to the differential equation of Fickian diffusion, in which the eddy diffusivities have been related to the standard deviations via Taylor's statistical theory; (b) it can be regarded as a purely empirical result embodying a bivariate normal distribution and continuity; (c) it can be seen as the result of some general statistical considerations in which the postulate of the

Gaussian distribution can be justified non-rigorously. Taylor's theorems and elementary statistical theory both entail the use of the ergodic principle whereby the ensemble mean concentration field at a given time is equivalent to the time-average concentration field observed in a single experiment, or alternatively, that time-average concentrations at fixed points can be related to particle displacement probabilities.

In any one derivation of the basic equation the explicit assumptions will vary according to the approach taken. A reasonably comprehensive list of all assumptions that may be involved is as follows:

- (a) The substance diffusing in the atmosphere behaves as if it consisted of ideal fluid "particles."

In effect the diffusing material is assumed identical in its physical properties with the idealized "fluid continuum", possessing neither extension, inertia, nor buoyancy of its own. Gases and submicron particles will remain suspended in the air for long periods of time and for practical purposes can be expected to behave in this manner. Low density particles up to 20 μm are also included.

- (b) The diffusion time is large compared to the Lagrangian time scale of atmospheric turbulence (of order 100 s).

For very small diffusion times the distribution of "particles" should take the same form as the wind fluctuation distribution since the "particle" trajectories coincide with the instantaneous wind. In the atmosphere this approximates a Gaussian distribution fairly closely.

- (c) The turbulence field is steady and homogeneous.

"Steady" or "stationary" means that the statistical properties derived from a time series do not vary with time. Thus a variance computed from a given hour of data would be the same as that computed from the next hour of data. Both diurnal and synoptic weather cycles cause major changes in the meteorological regime. The condition may be fulfilled approximately for intervals short compared with the diurnal cycle, that is, for a few hours (when the large-scale pattern does not change). "Homogeneous" means that the

statistical properties do not vary in space. A measurement at a particular point will display statistical characteristics identical to those taken at some neighboring point. Horizontal homogeneity is a reasonable assumption over regions of similar surface characteristics. Vertical homogeneity cannot, in general, be achieved because of the presence of the earth's surface and the action of the forces of buoyancy and gravity.

- (d) Diffusion occurs independently in the three coordinate directions and the distributions of the diffusing substance are jointly as well as separately normal.

The normal distribution for short diffusion times has already been justified. For longer times the essentially random nature of successive diffusion "steps" leads to the Gaussian distribution by the Central Limit Theorem. Detailed measurements have shown that even in very stable conditions the distributions are still separately normal, although in the extreme case they do not remain jointly normal (Gifford 1968).

- (e) Diffusion in the downwind direction is negligible compared to transport by the mean wind.

This assumption is not valid if the emission time is less than the travel time or if the mean wind tends to zero (say, less than 1 m/s).

- (f) The mean wind is uniform in space and time.

The surface boundary layer is characterized by marked vertical shear of the mean wind. However, horizontal homogeneity over uniform terrain at any fixed level is a reasonable assumption. The inadequacy can be compensated by the appropriate choice of an "effective" or "representative" mean wind. Atmospheric flows can never be regarded as stationary (steady) for long. A reasonable approximation might be expected for intervals short compared with the typical period of synoptic weather cycles (of order 100 h), but in reality such intervals occur infrequently due to the presence of intermediate-sized eddies. This is manifest as a "shift" or trend in the mean wind.

- (g) The surface of the earth is a perfect reflector and the equation of continuity is satisfied.

It is assumed that none of the material emitted is removed from the plume. Deposition, reaction, washout, and other removal processes are ignored. The presence of a solid boundary at ground level is represented by a "mirror image source" below ground. Experimental evidence supports this artifice as providing a satisfactory description of the concentration above the earth's surface. This assumption is modified when deposition is being estimated.

The Gaussian distribution yields small non-zero concentrations at large distances from the plume axis. This implies that diffusion proceeds in some sense with infinite velocity. Physically the speed of diffusion must be less than some finite value such as the speed of sound. A hyperbolic diffusion equation, the so-called telegrapher's equation, which allows for a limited propagation velocity, gives results that differ only slightly and are of little practical concern (Pasquill 1974, Gifford 1968).

The Gaussian plume diffusion model for a continuous elevated point source (assuming reflection at the ground) is represented by the equation

$$\chi(x,y,z;H) = \frac{Q}{u2\pi\sigma_y\sigma_z} \exp \frac{-y^2}{2\sigma_y^2} \exp - \frac{(h_{\text{eff}} - z)^2}{2\sigma_z^2} + \exp - \frac{(h_{\text{eff}} + z)^2}{2\sigma_z^2}$$

where

- χ is the time-averaged concentration of chemical;
- x,y,z are the distances downwind, cross-wind, and upward, respectively;
- h_{eff} is the effective source height above ground level (h_{eff} is equal to the sum of the physical stack height h_s and the plume rise Δh);
- Q is the source strength;
- σ_y is the standard deviation of the time-averaged plume concentration distribution in the cross-wind direction;
- σ_z is the standard deviation of the time-averaged plume concentration distribution in the vertical direction;
- u is the time-averaged wind speed

The factor before the exponential is the centreline concentration (in the absence of a solid boundary). The first exponential accounts for an off-axis location. The second exponential accounts for the elevation of the sources above the ground. The third exponential accounts for the ground surface, treating it as a perfect reflector with an image source.

Any consistent set of units may be used. It is customary to employ the SI system of units, so that $x, y, z, \sigma_y, \sigma_z$ are all in metres and u is in metres per second. The sources strength Q can be expressed as a mass emission rate (g/s) which yields a concentration as a mass per unit volume ($\mu\text{g}/\text{m}^3$) or as a volumetric flow rate (m^3/s) which yields a concentration as a ratio (fraction or ppm).

A variety of systems exist for specifying the sigmas (Weber 1976, Gifford 1976). The empiricism of these systems has prompted the assertion that "the great merit of the Gaussian plume approach.... is that it incorporates the available observational data on plume behavior to the maximum possible extent" (Gifford 1974b). Once the plume rise and standard deviations have been specified as functions of downwind distance, a variety of analytical solutions can be found for maxima (Ragland 1976) as was done by Kumar and Djurfors (1977) and functional behavior can be explored (Gifford 1976) as was done by Leung et al. (1972).

4.2 EMPIRICAL SYSTEMS FOR SPECIFYING PLUME SPREADS

These methods use graphical or power law summaries of experimental data and field measurements, guided by theoretical considerations.

4.2.1 The British Meteorological Office 1958 System

The spread estimates introduced by The British Meteorological Office (Meade 1960, Pasquill 1961) were introduced into North America as standard deviations (Gifford 1961). These now famous Pasquill-Gifford curves consisted of log-log plots of standard deviation as a function of distance and stability category (Figures 4.1 and 4.2). Although widely used, the basis and limitations of the estimates are often overlooked.

The curves were not just a compilation of observational experience; some a priori

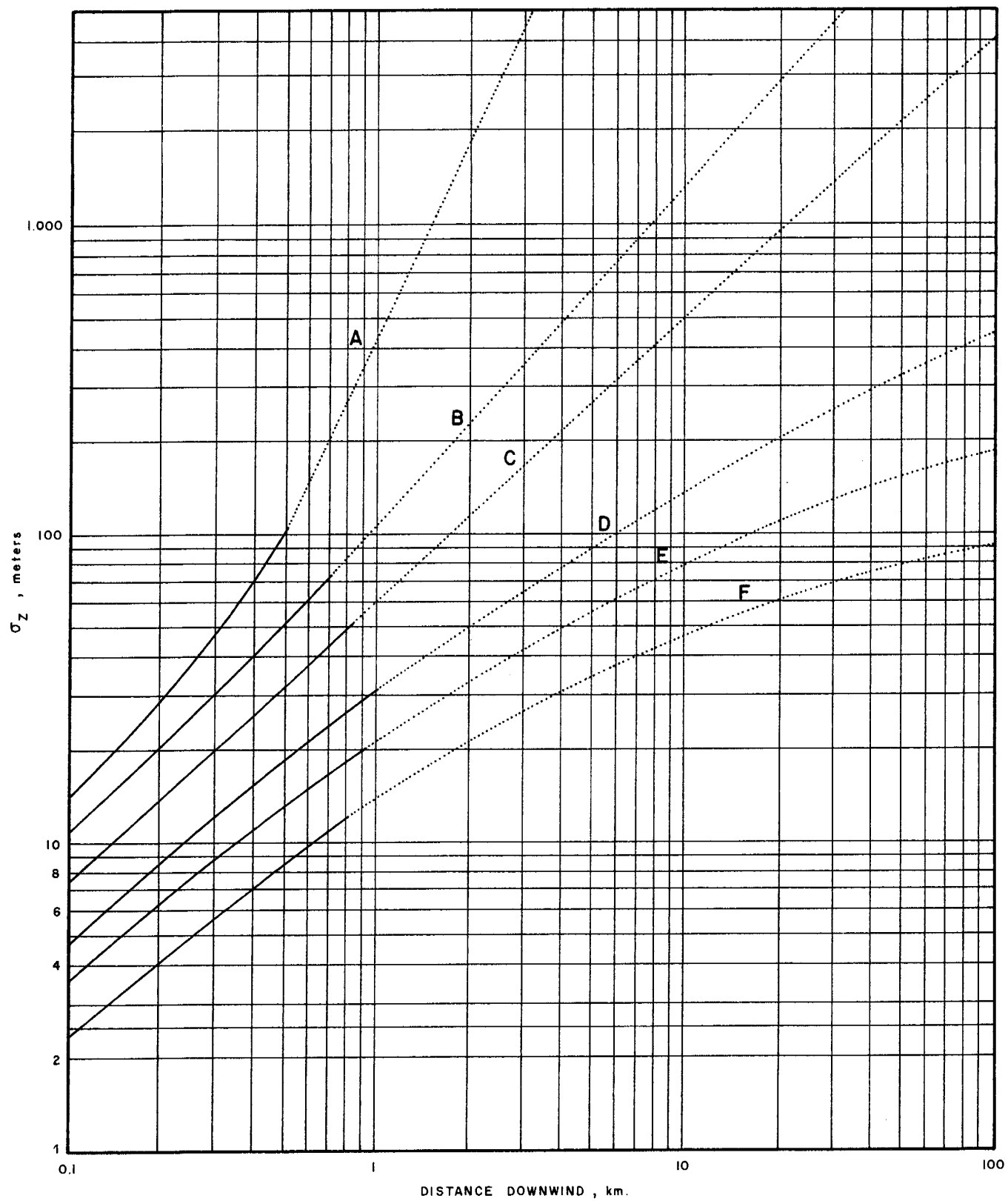


Figure 4.1 Vertical dispersion coefficient as a function of downwind distance from the source, after Pasquill and Gifford (Turner 1970).

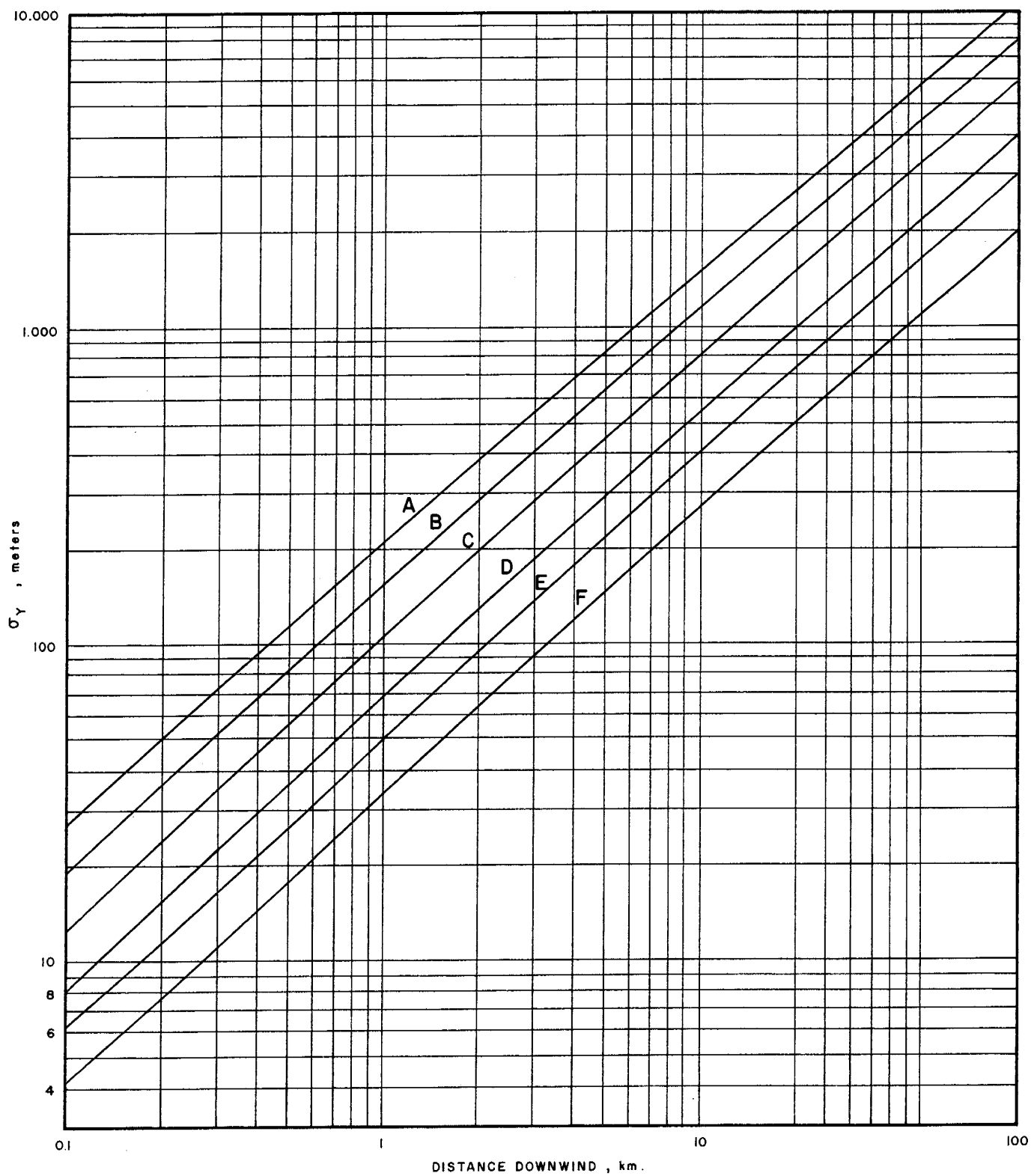


Figure 4.2 Horizontal dispersion coefficient as a function of downwind distance from the source, after Pasquill and Gifford (Turner 1970).

elements were embodied (Pasquill 1976). The cross-wind spread estimates for 0.1-1.0 km were derived by the method of Hay and Pasquill from wind fluctuation statistics over open downland (roughness 3 cm). The sampling time was 3 minutes. For the range 10-100 km the estimates were speculative, based on limited observations of tracer dispersion over level terrain of roughness about 30 cm.

For variable wind direction or longer period concentration averages (an hour or more) the values of lateral spread were to be based on the wind direction range, that is, the difference between the extreme values of wind direction during the period. For distances of up to a few kilometres the range over the averaging period was to be used as a multiplier of the value given by the graph; for distances greater than a few kilometres, the range of 15-minute average values of wind direction was to be used.

The vertical spread estimates for short distances (order of 1 km) were based on the Project Prairie Grass data, consolidated by Calder's gradient transfer results for a roughness of 3 cm. At longer range the estimates were based on limited experimental data and the few statistics available on vertical gustiness in the lower atmosphere. For the extreme cases of stability and instability the estimates were essentially speculative extrapolations of short-range data. In the original curves the confidence associated with the estimates is represented by the thickness and continuity of the lines.

The vertical sigmas referred strictly to sources at ground level, or at least sources such that $h_{\text{eff}} < \sigma_z$. In the absence of evidence or argument to the contrary, the estimates were also suggested for use with elevated sources.

In the original papers the sampling time implicit in the estimates was not made clear. This gap has since been filled (Pasquill 1974, 1976). For a surface release σ_z should become independent of sampling time beyond some limiting value which corresponds to the time required for the vertical distribution to respond to the whole spectrum of the vertical component of turbulence. Experience suggests that this limiting time will increase, at first linearly with height, and then more slowly. As a rough guide the estimates should apply to a minimum sampling time of 10 minutes for $h_{\text{eff}} \geq 100$ m and to times of $(h_{\text{eff}}/10)$ minutes for $h_{\text{eff}} < 100$ m.

These essential points are summarized in Table 4.2. It was emphasized that the curves were essentially tentative and that they required progressive confirmation or adjustment as more information became available.

To facilitate computations, it has been found useful to fit power laws to the Pasquill curves (Tadmor and Gur 1969, Martin and Tikvart 1968, Eimutis and Konicek 1972). The obvious limitation is that no one power law can fit the vertical spread data over all distances. Consequently most workers have used two or three different sets of power law constants to match the sigma values over various ranges of distances. On the log-log graph paper this appears as a segmented line with each segment having a different slope. Single power law approximations (Ragland 1976, van der Hoven 1972) are given in Table 4.3.

Johnson (1967) reported that the Pasquill scheme gave reasonable agreement with the phosgene diffusion trials carried out at Suffield in 1972. Ratios of predicted to observed crosswind integrated dose ranged from 0.6 to 1.3 while ratios of predicted to observed aerial dosage ranged from 0.85 to 1.55 for distances up to 40 km.

Hollenbach (1960) stated that the method gave satisfactory correlations with measured SO_2 concentrations in the Joffre D-2 field.

Lee et al. (1973) found that under stable conditions, the plume from the Strachan gas plant had a horizontal spread equivalent to category C while the vertical spread was the indicated F. Limited mixing conditions with a 600-m mixing layer produced a horizontal spread larger than class A and a vertical spread invariant with distance. Under neutral conditions at both Rimbey and Nevis the horizontal spreads were equivalent to the class A curve and vertical spreads, class C and B respectively. The enhanced vertical spread was attributed to deflection by the steep river bank.

Fanaki (1978) found that the vertical spread of the Suncor plume near Ft. McMurray was larger than the values from the Pasquill curves. During stable conditions, the values were closer to the neutral curves. All values were measured within 1 km of the stack, so clearly they were dominated by the buoyancy-induced spreading which dominates the plume rise region of dispersion. Fanaki et al. (1979b) reported that horizontal spread measurements at distances out to 20 km were also larger than the

Table 4.2 Summary of the basis and scope of the original Pasquill-Gifford curves (Pasquill 1976).

<u>Crosswind Spread</u>	
Source height h_{eff}	any (within mixed layer)
Sampling Time τ	3 minutes
Basis for $x = 0.1 - 1$ km	preliminary statistics of wind direction fluctuation for $z_0 \approx 3$ cm
Basis for $x = 10 - 100$ km	extrapolation of short-range data in the light of limited special observations of tracer dispersion over level terrain of mixed roughness (implied $z_0 \approx 30$ cm)

<u>Vertical Spread</u>	
Source height h_{eff}	effectively zero, but offered as usable for any height in a mixed layer in the absence of strong evidence against.
Sampling time τ	any; for an elevated source the estimates of σ_z are appropriate to sampling times in excess of some limiting value τ_m , roughly proportional to height of source h_{eff} for h_{eff} up to about 100 m, at which height and above τ_m is roughly 10 minutes.
Basis for $x = 0.1 - 1$ km	properties of the wind profile over a surface of small roughness ($z_0 = 3$ cm), with guidance from diffusion studies as to the effect of thermal stratification.
Basis for $x = 10 - 100$ km	as for σ_y , with guidance from early data on the properties of the vertical component of turbulence at heights throughout the mixed layer.

Table 4.3 Single power law approximations to the Pasquill-Gifford curves.

(a) Ragland (1976)

$$\sigma_z = ax^b \quad \sigma_y = cx^d \quad 1 \text{ km} < x < 100 \text{ km}$$

$\sigma, x \text{ in m}$

Stability	a	b	c	d
Very unstable (A)	0.00022	2.10	0.59	0.85
Moderately unstable (B)	0.056	1.10	0.41	0.86
Slightly unstable (C)	0.12	0.91	0.24	0.88
Neutral (D)	0.73	0.55	0.14	0.89
Slightly stable (E)	0.82	0.48	0.11	0.89
Moderately stable (F)	0.63	0.45	0.075	0.89

(b) van der Hoven (1972)

$$\sigma_z = 1000 ax^b \quad \sigma, x \text{ in km}$$

$$\sigma_y = 1000 cx^d \quad \sigma \text{ in m}$$

Stability	a	b	c	d
Very unstable (A)	0.45	2.1	0.20	0.88
Moderately unstable (B)	0.11	1.1	0.16	0.88
Slightly unstable (C)	0.061	0.92	0.10	0.88
Neutral (D)	0.033	0.60	0.070	0.88
Slightly stable (E)	0.025	0.51	0.052	0.88
Moderately stable (F)	0.015	0.45	0.035	0.88

$$2 \text{ km} \leq x \leq 70 \text{ km}$$

Pasquill values. This was attributed to the trapping of the plume beneath an elevated inversion. Fanaki et al. (1979a) found that observed vertical spread values in stable conditions did not increase with distances. Enhanced horizontal spreads were attributed to valley wind shear.

Davison et al. (1976) used an instrumented aircraft to measure the dispersion of the Suncor plume. The horizontal spread values were much larger than those indicated by the Pasquill curves, but their variation with downwind distance was similar. In contrast, the vertical spreads did not increase as quickly with distance, which was attributed to the presence of elevated inversions. The enhanced horizontal spread was attributed primarily to the presence of multiple sources, but also to topographically induced turbulence which also enlarged the vertical spread at a distance of 5 km.

Whaley and Lee (1978) used an instrumented helicopter to measure plume spreads at the same Suncor plant in northeastern Alberta. They similarly observed that the sigmas were larger than Pasquill's values by virtue of initial buoyancy and terrain effects. In neutral conditions at 10 km the product was a factor of 5 larger with little difference between winter and autumn values. In stable conditions at 20 km the product was comparable to Pasquill's F curve, but at closer distances winter values were substantially larger than fall values. Whaley et al. (1980) reported quite disparate results for two sour gas plants in southwest Alberta. The horizontal sigma under neutral conditions for Pincher Creek was wider than Pasquill's A values, but for Waterton was narrower than Pasquill F. Vertical sigmas were between A and C values for both plumes. The lack of spread of the Waterton plume was attributed to the strongly channelled flow induced by mountain passes immediately upwind. The enhanced vertical spread was attributed to the turbulence caused by the mountains. Whaley and Lee (1982) summarized 6 years of plume measurements across Canada and concluded that the horizontal spreads were always larger by 2 - 6 classes and that the vertical spread agreed between 4 and 10 km, closer to the stack, σ_z was larger and further out σ_z was smaller.

Slawson et al. (1978) measured sigma values for the Suncor plume that were broadly consistent with the Pasquill values; however when classified by bulk Richardson number, the slopes of the curves were different, steeper in neutral/unstable and shallower

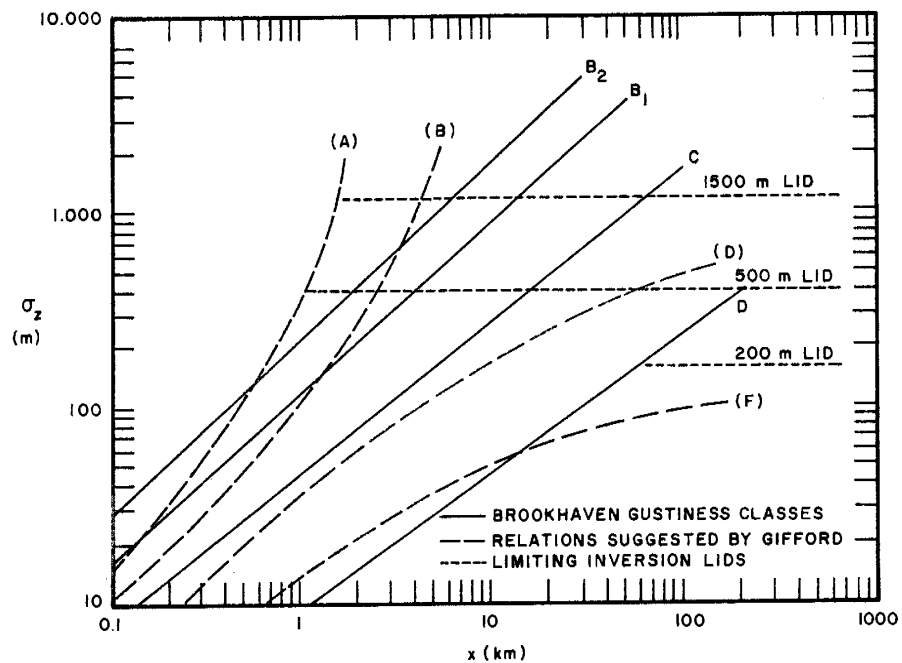
in stable conditions. Cheng et al. (1987) reported that their measurements of the Syncrude plume gave slopes consistent with those of the Pasquill curves, but that the magnitudes were larger than would be indicated for the stability class.

4.2.2 Brookhaven National Laboratory (BNL)

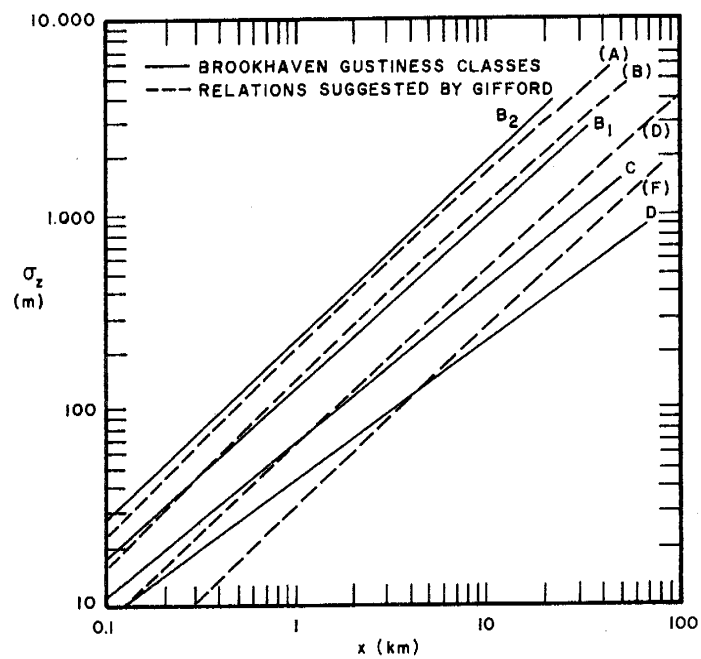
Brookhaven National Laboratory developed a set of dispersion curves (Figure 4.3) and equations (Table 4.4) based on a variety of data taken over a 15-year period (Singer and Smith 1966). The BNL site is slightly rolling terrain with a mixture of metre-high grass and trees 6 - 8 m high (DeMarrais 1959) with an average roughness of about 100 cm. Most of the data were from oil fog released at a height of 108 m. Three types of measurements were made: (a) photometric densitometers were used to record the concentration variations; (b) the fluorescence of the oil was used to obtain hourly mean samples; (c) the dimensions of the visible plume provided estimates of some parameters. Radioactive argon emitted from 108 m allowed the determination of parameters at distances of 50 km and more. Uranine dye from a 2-m source also provided a small amount of information at short range. The cross-wind spreads were measured directly, while the vertical spread was calculated from the measurements of horizontal spread, concentration and wind speed. The concentration data were mean values obtained over periods ranging from 30 to 90 minutes, that is, on the order of one hour. The data were presented in terms of gustiness classes which, according to the authors, "seem to suggest that the atmosphere may have preferred dispersion conditions."

The curves themselves exhibit two significant differences from the foregoing Pasquill-Gifford curves (Figure 4.3).

- (a) There is no curvature in the σ_z plot. There are two possible reasons for this (Singer and Smith 1966). In stable cases, σ_z is a very gentle function of distance and only with precise data can the difference between a straight line and a curve be determined. In unstable cases the explanation may lie with the method of calculation. The average wind speed u is usually assumed constant, even though it must increase as the vertical depth of the cloud increases. If it is held constant while it is actually increasing the σ_z plot will curve upward to compensate. For ground-level sources and 10-m winds the tendency would be pronounced. This may account for the upward curvature



(a) Vertical spread.



(b) Cross-wind spread.

Figure 4.3 The dispersion curves of the Brookhaven National Laboratory compared with the Pasquill-Gifford curves (Smith and Singer 1966).

Table 4.4 BNL plume standard deviations as a function of distance and meteorological conditions (Smith and Singer 1966).

Gustiness Class	σ_y	σ_z	σ_z/σ_y
B ₂	$0.40 x^{0.91}$	$0.41 x^{0.91}$	1.03
B ₁	$0.36 x^{0.86}$	$0.33 x^{0.86}$	0.91
C	$0.32 x^{0.78}$	$0.22 x^{0.78}$	0.70
D	$0.31 x^{0.71}$	$<0.06 x^{0.71}$	<0.20

Table 4.5 TVA diffusion coefficients (Ragland 1976).

$$\sigma_z = ax^b, \sigma_y = cx^d \quad 1 < x < 100 \text{ km}$$

$\sigma, x \text{ in m}$

Stability	a	b	c	d
Neutral	0.37	0.74	0.37	0.76
Slightly stable	1.42	0.50	0.55	0.69
Stable	2.94	0.34	0.78	0.63

of the Pasquill-Gifford curves. The Brookhaven curves, however, were obtained with a wind speed measured at the source height of 108 m and since the plume disperses both upward and downward, the elevated reference wind speed is valid for a considerable distance. The downward curvature of the Pasquill-Gifford neutral and stable curves could have been due to the presence of an elevated inversion lid (Smith and Singer 1966).

- (b) The σ_y and σ_z variations with downwind distance are identical, that is, the power law exponents are equal. This greatly simplifies many calculations and may have contributed to the decision of the American Society of Mechanical Engineers to adopt these coefficients in their guidebook (Smith 1968).

Lidar measurements of the Suncor plume by Fanaki et al. (1979b) gave spread values in reasonable agreement with the ASME (BNL) curves for unstable and neutral atmospheric conditions. Under stable conditions, the spreads were larger than indicated by the curves and did not grow with distance, a discrepancy attributed to buoyancy effects. Davison and Leavitt (1979) found that the BNL curves for horizontal spread fit the oil sands data reasonably well, but the vertical spread curves did not.

4.2.3 Tennessee Valley Authority (TVA)

The Tennessee Valley Authority has summarized 20 years of experience with the measurement of SO_2 concentration patterns around power plants with stacks ranging in height from 75 to 250 m (Carpenter et al. 1971). All plume spread data were obtained by means of an instrumented helicopter flying stacked traverses at different downwind distances. The quasi-instantaneous samples were obtained within the 2 - 5 minutes that it took the helicopter to pass through the plume. The data and resulting curves were categorized by the potential temperature gradient at plume height. Plume heights were rarely less than twice the physical stack height. Since superadiabatic lapse rates do not normally occur at such altitudes, the TVA categorization contains no unstable classes whatsoever.

At short range there was little separation of the sigma values by stability class, but at long range the more neutral classes exhibited a slight upward curvature. The growth rate was less than the Pasquill-Gifford curves, that is, the slope of the lines on the

log-log plot is less. At short range the magnitudes were substantially larger than those of Pasquill-Gifford. Since buoyant plumes were involved this may well have been the effect of buoyancy-induced initial plume expansion.

A power law fit to the TVA curves (Ragland 1976) is given in Table 4.5. Fanaki et al. (1979a) reported that vertical sigmas measured on the Suncor plume followed the TVA curves more closely than the Pasquill curves. Davison and Leavitt (1979) examined all the oil sands sigma data and found that the TVA scheme worked well for the vertical sigma under stable conditions.

4.2.4 Briggs Formulas

A set of interpolation formulas for the data contained in the aforementioned empirical systems has been proposed (Briggs 1973) as in Table 4.6. The PG curves were developed using diffusion data out to a distance of about 1 km and a passive (non-buoyant) tracer. The BNL curves also reflected non-buoyant plume dispersion data, but from an elevated source and most measurements were made at distances of a few kilometres. The TVA curves were based on data from still greater effective heights (150 - 600 m or more) and downwind distances of up to tens of kilometres. The short-range data of TVA were suspected to reflect enhanced spreading rates due to buoyant entrainment; however, at greater distances atmospheric turbulence properties ought to be reflected. Other than at small distances (where the TVA curves display strong plume buoyancy effects) the TVA and BNL curves agree reasonably well with one another and, except for A and B conditions, with the PG curves at about 10 km. Beyond 10 km the TVA curves are less steeply inclined.

By weighing the data for the appropriate distance range Briggs prepared a composite set of curves that followed the PG curves fairly closely out to 10 km (Figure 4.4) except for classes A and B which followed the BNL curves beyond 1 km. For distances beyond 10 km his curves resembled the TVA's. The curves for σ_y were described by mathematical formulas which also satisfied the statistical theory prediction of growth, namely, $\sigma_y \propto x$ at short distances and $\sigma_y \propto x^{1/2}$ at long range. Briggs also included a family of urban dispersion curves based on the McElroy and Pooler (1968) data. The formulas were to be used only for the estimation of ground-level maximum concentrations, not for axial plume concentrations. Thus, greater distances reflect greater source heights.

Table 4.6 Briggs plume spread formulas for elevated small releases,
 $0.1 \leq x \leq 10$ km (Gifford 1975).

	Pasquill Type	σ_y (m)	σ_z (m)
Open Country	A	$0.22x(1+0.0001x)^{-1/2}$	$0.20x$
	B	$0.16x(1+0.0001x)^{-1/2}$	$0.12x$
	C	$0.11x(1+0.0001x)^{-1/2}$	$0.08x(1+0.0002x)^{-1/2}$
	D	$0.08x(1+0.0001x)^{-1/2}$	$0.06x(1+0.0015x)^{-1/2}$
	E	$0.06x(1+0.0001x)^{-1/2}$	$0.03x(1+0.0003x)^{-1}$
	F	$0.04x(1+0.0001x)^{-1/2}$	$0.016x(1+0.0003x)^{-1}$
Urban Areas	A-B	$0.32x(1+0.0004x)^{-1/2}$	$0.24x(1+0.001x)^{1/2}$
	C	$0.22x(1+0.0004x)^{-1/2}$	$0.20x$
	D	$0.16x(1+0.0004x)^{-1/2}$	$0.14x(1+0.0003x)^{-1/2}$
	E-F	$0.11x(1+0.0004x)^{-1/2}$	$0.08x(1+0.0015x)^{-1/2}$

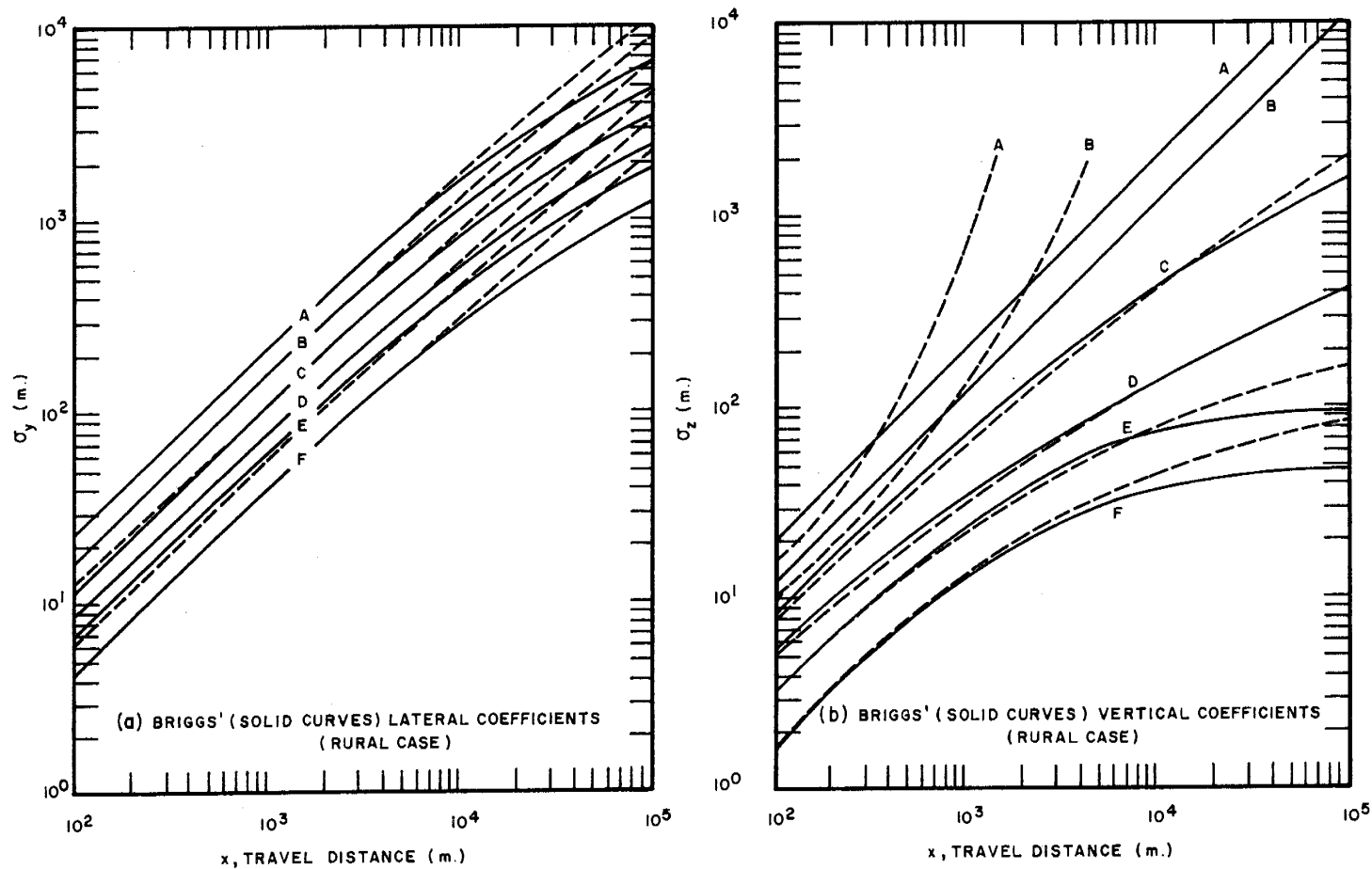


Figure 4.4 Briggs' rural dispersion coefficients compared to the Pasquill-Gifford coefficients (Hosker 1973).

Davison and Leavitt (1979) found that the oil sands sigma data were no better represented by the Briggs curves than by the Pasquill curves.

4.3 PLUME SPREADS FROM TAYLOR STATISTICAL THEORY

An expression relating particle displacements for a continuous source and Lagrangian velocity fluctuations (assuming homogeneity and stationarity) was first derived more than 70 years ago (Taylor 1921):

$$\sigma_y^2 = 2\sigma_v^2 \int_0^T \int_0^t R(\xi) d\xi dt$$

where T is the dispersion time, R is the Lagrangian autocorrelation and σ_v^2 is the variance of the wind component fluctuations. At short dispersion times, R is approximately 1 so

$$\sigma_y^2 = \sigma_v^2 T^2$$

At very long dispersion times,

$$\sigma_y^2 = 2\sigma_v^2 t_L T$$

where t_L is the Lagrangian time scale of turbulence.

Thus, Taylor's theorem says that near the source the plume spread is proportional to distance ($x = uT$), while at large distances it is proportional to the square root of distance. This change in exponent arises because larger and larger scales of motion contribute to the spread and smaller scales become progressively less important. Although neither the form of the autocorrelation function, nor the limiting times are known, the results are not sensitive to wide variations in the functional form (Pasquill 1974).

$$\text{The relationship can also be written as } \sigma_y = \sigma_v T f_1(T/t_L)$$

where f_1 is a universal function of dispersion time T and Lagrangian time scale t_L which has a small T limit of $f_1 = 1$ and a large T limit of $2 t_L^{1/2}/T$ (Pasquill 1971, 1974).

4.3.1 Hay - Pasquill

The equivalent Lagrangian averaging method (Hay and Pasquill 1959) assumes that Lagrangian statistics can be replaced by Eulerian (fixed-point) statistics taken over an appropriately shortened time interval. According to this method the plume spread can be predicted irrespective of thermal stratification by the relation

$$\sigma_y = x \sigma_\theta (\tau, x/\beta u)$$

where

- σ_θ is the standard deviation of the wind direction in radians
- τ is the sampling period
- $x/\beta u$ is the length of the running average to be used in smoothing the time series
- β is the ratio between Lagrangian and Eulerian time scales
- x is the downwind distance
- u is the mean wind speed

A similar expression applies for σ_z in terms of the elevation angle fluctuation σ_ϕ . Although a value of 4 has generally been used for β , it is more accurately given by $\beta = 0.44/i$ where i is the intensity of turbulence (Pasquill 1974). This moving-average variance method (Gifford 1968) has several advantages: (a) it specifies diffusion coefficients by a detailed analysis of atmospheric turbulence measurements; (b) it does not involve adjustable empirical constants; (c) it is comparatively free from debatable physical assumptions; (d) it is not in principle limited to a particular range of diffusion times. However, it is limited to the conditions under which Taylor's theorem applies. The main practical problem is that the required wind fluctuation data are rarely available, although the advent of microcomputer data loggers is changing this.

Walker (1965) reported that good agreement was obtained between standard deviations measured in particulate diffusion experiments at Suffield and values predicted by this method under stationary conditions. The ratio between Lagrangian and Eulerian time scales was found to increase as turbulence intensity decreased. Results were improved using measured β values as opposed to the standard value of 4.

The success of the Hay-Pasquill formulation in real-world situations indicates that it is "more important to allow for the actual, nonstationary character of atmospheric turbulence, than to have accurate information on the Lagrangian correlation coefficient" (Csanady 1973).

4.3.2 Draxler

Many historical sets of diffusion data have been reanalyzed (Draxler 1976) to determine the form of $f(T/t_L)$. Since the Lagrangian time scale could not be determined from most data sets, a quantity T_i proportional to t_L is defined such that for $T \rightarrow 0$, $f(T_i) = 1.0$ and for $T = T_i$, $f(T_i) = 0.5$. For each experimental data, set graphs of f versus $1 + T/T_i$ were plotted for horizontal and vertical spreads. Significant differences were found between ground-level releases and elevated sources. His best-fit equations are shown in Table 5.7. The parabolic curve for vertical diffusion from a ground-level source in unstable conditions was an approximation to widely scattered data points. The vertical diffusion from an elevated source in stable conditions drops off more quickly than in other conditions. Neither of these two equations satisfies Taylor's limits.

The Pasquill-Gifford curves consistently underestimated the horizontal spread of a plume from an elevated source. Draxler's equations and those based on an experimental autocorrelation function were shown to fit the data better. Horizontal spreads in the oil sands area were predicted reasonably well, with Davison and Leavitt (1979) reporting 75% of values within a factor of 2.

4.3.3 Other Universal Functions

Alternative functional forms (Cramer 1976, Pasquill 1976, Irwin 1983 and Gryning et al. 1987) are shown in Table 4.8. Several of these alternatives have been compared against a large body of tracer data (Irwin 1983). For the lateral spread of near-surface releases, Pasquill's formulation was better, while for elevated releases, Cramer's was superior. Draxler's scheme was almost as good in both cases. Draxler's scheme was also better for vertical spread although Cramer's also worked well for unstable conditions. Choosing forms invariant with height resulted in little overall loss in performance. The Pasquill scheme underestimated horizontal spreads in the oil sands area by about 25% according to Davison and Leavitt (1979).

Table 4.7 Summary of Draxler's equations.

Stability		Horizontal Spread		Vertical Spread	
		Stable	Unstable	Stable	Unstable
ground sources					
f	(T<550)	$\frac{1}{1+.9(t/T_i)^{0.5}}$	$\frac{1}{1+.9(t/T_i)^{0.5}}$	$\frac{1}{1+.9(t/T_i)^{0.5}}$	$\frac{0.3(t/T_i-0.4)^2}{0.16} + .7$
f	(T>550)	$\frac{1}{1+28/t^{0.5}}$			
T _i		300	300	50	100
Elevated Sources					
f		$\frac{1}{1+.9(t/T_i)^{0.5}}$	$\frac{1}{1+.9(t/T_i)^{0.5}}$	$\frac{1}{1+.945(t/T_i)^{0.806}}$	$\frac{1}{1+.9(t/T_i)^{0.5}}$
T _i		1000	1000	100	500

Table 4.8 Other universal functions.

Author	Lateral Diffusion f_y	Vertical Diffusion f_z
Cramer (1976)	$\frac{(50)}{x} \left\{ \frac{x-5}{45} \right\}^{0.9}$	1.0
Pasquill (1976)	for $x < 10,000$ $\frac{1}{1 + (x/2500)^{0.5}}$	---
	for $x > 10,000$ $\frac{1}{3 (x/10,000)^{0.5}}$	---
Irwin (1983)	$\frac{1}{1 + 0.9 (T/1000)^{0.5}}$	unstable $\frac{1}{1 + 0.9(T/500)^{0.5}}$ stable $\frac{1}{1 + 0.9(T/50)^{0.5}}$
Venkatram et al. (1984)	----	stable $\frac{1}{(1 + T/2T_L)^{0.5}}$ where $T_L = \ell/\sigma_w$ $1/\ell = 1/36z + N/(.27 \sigma_w)$ $N = \left\{ \frac{g}{T} \frac{d\theta}{dz} \right\}^{1/2}$
Gryning et al. (1987)	elevated sources $\frac{1}{(1 + T/1200)^{0.5}}$ ground sources $\frac{1}{(1 + T/400)^{0.5}}$	slightly unstable $\frac{1}{(1 + T/600)^{0.5}}$ stable $\frac{1}{(1 + T/60)^{0.5}}$

4.4 PLUME SPREADS FROM INTERRELATED THEORIES

Gradient transfer equations and boundary layer scaling relations can be used to estimate diffusion. The results can then be converted into the equivalent Gaussian sigma.

4.4.1 F.B. Smith Curves

The two-dimensional gradient transfer relations can be solved numerically using K values specified in terms of measurable turbulence properties and which accordingly contain the effects of surface roughness and heat flux (Smith 1972, Pasquill and Smith 1983). The solutions were used to construct nomograms giving σ_z in terms of stability (using an objective stability classification based on wind speed and sensible heat flux with an equivalence with the old Pasquill system) and surface roughness. For simplicity, power law approximations were provided (Table 4.9) over distances 0.1 - 10 km.

For neutral flow over terrain of roughness 10 cm the revised values were similar to the original Pasquill values. The main difference occurred in the unstable categories, where the pronounced acceleration of growth was no longer present (Figure 4.5).

The extreme effect of the combination of roughness and heat flux is to produce a range of about 30:1 in σ_z . At short downwind distances, roughness and heat flux have similar importance, but as distance increases the importance of roughness decreases while that of heat flux increases (Pasquill 1975).

The specification of σ_z was considered tentative at distances greater than a few kilometres because of uncertainties in the estimates of K in the upper part of the boundary layer, especially in unstable conditions when buoyancy forces dominate frictional forces.

Davison and Leavitt (1979) found that the Smith curves overestimated the observed vertical spreads in oil sands plumes. However, this was attributed to the incorrect selection of stability category over snow surfaces.

Table 4.9 Power law approximations for the growth of vertical spread with distance for a source at ground level according to F. B. Smith (Pasquill and Smith 1983).

Stability category	z_0	Coefficient a			Index b		
		1 cm	10 cm	1 m	1 cm	10 cm	1 m
A		0.102	0.140	0.190	0.94	0.90	0.83
B		0.062	0.080	0.110	0.89	0.85	0.77
C		0.043	0.056	0.077	0.85	0.80	0.72
D		0.029	0.038	0.050	0.81	0.76	0.68
E		0.017	0.023	0.031	0.78	0.73	0.65
F		0.009	0.012	0.017	0.72	0.67	0.58

The figures refer to the form $\sigma_z = ax^b$, with σ and x in km, fitted at $x = 0.3$ and 3 to estimates from nomograms. These estimates actually lie on log-log plots which are concave to the log x axis, and with respect to these curves the power law fittings generally underestimate σ_z at $x = 1$ km by roughly 5 percent and overestimate at $x = 0.1$ or 10 km by roughly 10 percent. It is emphasized that the error in this representation of the nomogram increases rapidly beyond $x = 10$ km.

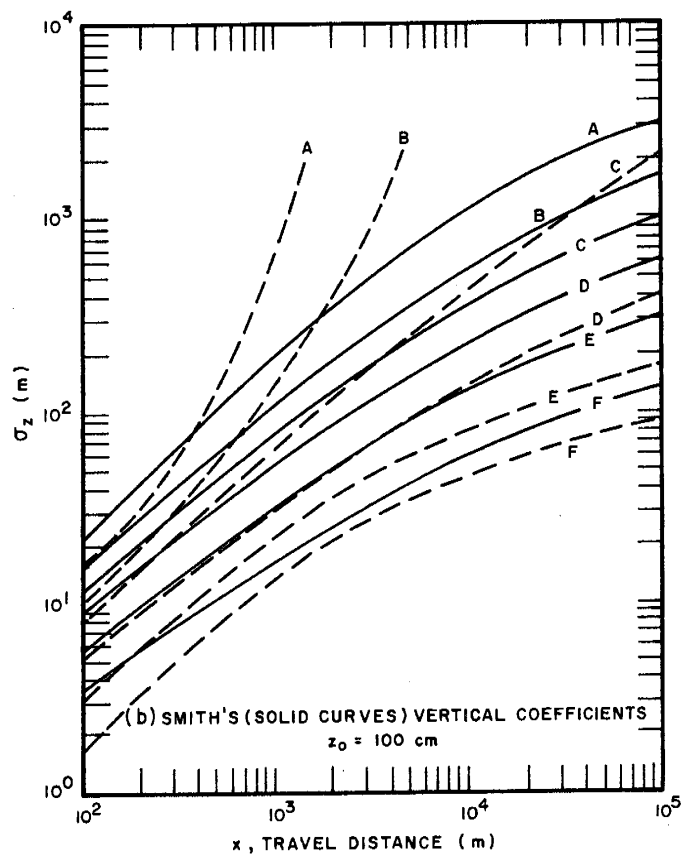
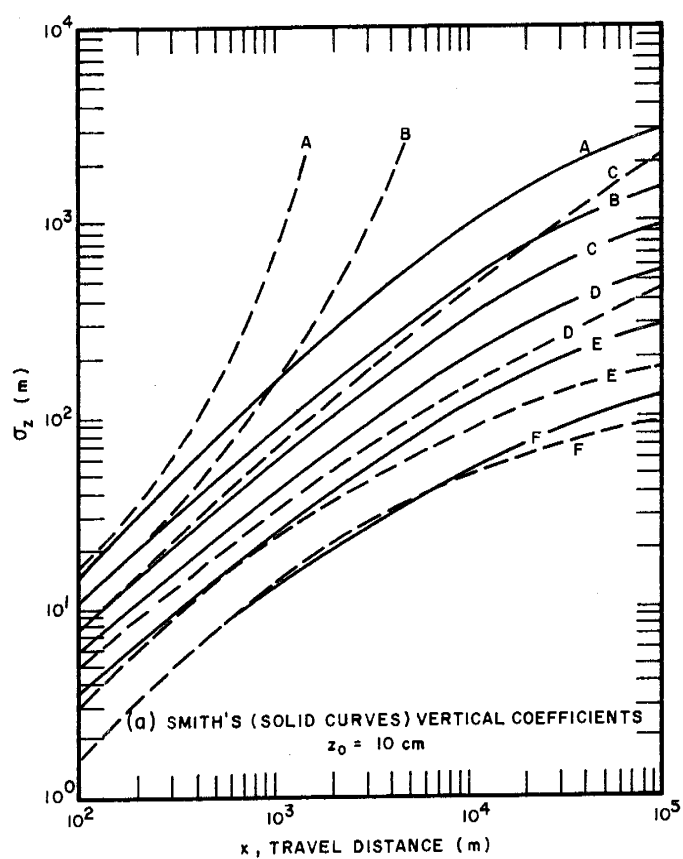


Figure 4.5 Smith's $\sigma_z(x)$ compared to Pasquill-Gifford $\sigma_z(x)$ for all stability categories and two roughness lengths (Hosker 1973).

Similarity theory can be extended beyond the surface layer using as parameters the standard deviation of the vertical wind and the length scale of turbulence (Pasquill 1976). Both were regarded as functions of height. The results were consistent with F. B. Smith's solutions, but data in suitable form were not available for observational testing. A useful consequence of the development is the capability of adjusting the F. B. Smith solutions for different K profiles without recourse to the numerical solution of the diffusion equations.

This local similarity approach has been used to examine the vertical spread in unstable conditions (Pasquill 1976). A free convection law of the Deardorff type gives results similar to those of F. B. Smith for PG category A over the height range for which σ_w follows the free convection similarity law. If the law is applied to much greater heights, the values approach the PG estimates but are still on the low side.

Several curves of σ_z versus x have been produced based on the second-order closure approach (Lewellen and Teske 1975). Agreement with F. B. Smith's values is good in neutral conditions. "This approach offers, what seems to be at the present, the only hope of theoretically determining σ_z for intermediate range for elevated sources ($\sigma_z = H$) at which range the statistical and gradient transfer theory approaches are most questionable" (Weber 1976).

4.4.2 Boundary Layer Scaling

For unstable conditions numerous authors have evaluated plume spreads in terms of the convective scaling parameters w_* and h (Weil and Brower 1984, Weil 1985, Briggs 1985, Sakiyama and Davis 1988). To bridge between neutral and convective limits various interpolations have been proposed as shown in Table 4.10.

For neutral and stable conditions, most authors use the universal function formulation of Taylor's theorem and compute the standard deviation of the wind component fluctuations using the scale relations (Gryning et al. 1987).

Table 4.10 Sigma values from neutral to unstable using convective scaling.

AUTHOR	CROSS-WIND	VERTICAL
Weil and Brower (1984)	$\sigma_y = [0.08^2 \times (0.56 \frac{w_*}{u})^2] x$	$\sigma_z = [0.06^2 \times (0.56 \frac{w_*}{u})^2]^{1/2} x$
Weil and Brower (1982)	$\sigma_y = \frac{[(0.56 w_*)^2 + (1.6 u_*)^2]^{1/2} (x/u)}{(1 + x/10 \text{ km})^{1/2}}$	$\sigma_z = [(0.56)^2 + (1.26 u_*/w_*)^2]^{1/2} hX$
Berkowicz et al. (1986)	$\sigma_y = \left(\frac{0.25 w_*^2}{1 + 0.9X} + u^2 \right)^{1/2} \frac{x}{u}$	$\sigma_z = (\sigma_{zm}^2 + \sigma_{zc}^2)^{1/2}$ $\sigma_{zc}^2 = 0.33 w_*^2 t^2$ $\sigma_{zm}^2 = 1.2 u_*^2 t^2 \exp(-0.6 tu_*/h_{\text{eff}})$
where		
$X = \frac{x w_*}{u h}$		
$w_* = \left(\frac{gH h}{\rho c_p} \right)^{1/3}$		
h is the height of the convective layer		
H is the heat flux		
h _{eff} is the plume height		

Alp et al. (1990) reported that the plume spreads calculated this way give similar values to those of F. B. Smith for vertical spread and Pasquill for horizontal spread. They also found that the Weil and Brower formulation for convective conditions gave larger values of vertical spread than F. B. Smith and smaller values of horizontal spread than Pasquill.

4.5 AVERAGING TIME

The term "averaging time" refers to the arbitrary observation time used to collect field measurements of turbulence or concentration. Typically this time will be 3 minutes or 1 h. However, the term is sometimes used to designate the interval over which a fluctuating wind component is averaged prior to computation of the variance, and the observation time is then referred to as the "sampling duration" (Pasquill 1962, 1974). To avoid confusion, the fluctuation pre-averaging interval is called the "smoothing time". The term "averaging time" is then interchangeable with "observation time" and "sampling time".

4.5.1 Relative and Total Diffusion

Two phenomena are responsible for the dilution observed at any fixed point on the ground. Eddies of size similar to the plume will cause it to grow with respect to its centre of mass. This is called relative diffusion or two-particle diffusion because the theory is concerned with the separation of two-particles initially released together. Turbulent eddies larger than the plume will cause the entire plume to move sideways like a flag flapping in the breeze. The observer at a fixed point on the ground will measure concentrations at various times in the plume centreline, out of the plume altogether, or somewhere in between. As a result, the concentrations are reduced from those that could be measured by an observer moving with the plume. This total diffusion is called Eulerian diffusion, time-averaged diffusion or apparent diffusion. The component due to the "flapping" of the plume around its mean position is often called meandering, but such usage is inconsistent with the river analogy since the term implies a slow, winding motion. In the strictest sense a meandering plume would only be observed in a stable atmosphere when low frequency changes in wind direction are present. This type of diffusion is also known as one-particle diffusion since it determines the spread between single particles released serially from a given point.

A plume is considered to be from a continuous source if the emission time is much greater than the sum of travel time and averaging time. A plume is considered to be a puff or cluster if the emission time is much less than the observation time (Ide 1971). The diffusion of a continuous plume observed at a given averaging time is the same as the diffusion of a cluster emitted over an equal time but observed over a long interval. This reciprocity of observation and emission time means that the spread of particles released sequentially from a fixed point over some time interval is equivalent to the spread from a permanent continuous source when observed for the same time.

A plume can be classified according to observation (averaging) time (Figure 4.6): instantaneous if the averaging (observation) time is much less than the travel time; steady if the averaging time is much greater than the travel time; and quasi-steady if the averaging time is about the same as the travel time (Ide 1971). The spread rates for each type will fall into one of three regimes: (1) a linear zone where sigma is proportional to travel time (the short time limit of Taylor's theorem), (2) a "parabolic" zone where sigma tends to be proportional to the square root of travel time (the long time limit of Taylor's theorem), or (3) a cuspidal zone where sigma tends to be proportional to the 1.5 power of travel time in accordance with two-particle or cluster spreading theory. The steady plume consists of a linear zone near the source and a parabolic zone far downwind. The instantaneous plume consists of a cuspidal zone near the source and a parabolic zone far downwind. The quasi-steady plume consists of a linear zone, a cuspidal zone and a parabolic zone (Figure 4.7). The cuspidal zone occurs for travel times much greater than the averaging time but less than the Lagrangian integral time scale. This behavior is summarized in Table 4.11.

When a plume is traversed with a chemical monitor mounted in an aircraft or automobile, a measurement of relative diffusion is obtained, that is, the spread of the instantaneous plume. Davison and Leavitt (1979) argued that speed of traverse is largely irrelevant because no Eulerian (fixed reference frame) averaging is being done. A slow traverse simply generates scatter about the true instantaneous plume. Similarly the response time of the monitor does not represent time averaging but rather space averaging that simply distorts the cross-sectional shape. When repeated traverses are made, both relative and total spread can be estimated. If the traverse profiles are averaged with respect to the centre of mass (overlapping the central peaks) then an estimate of the

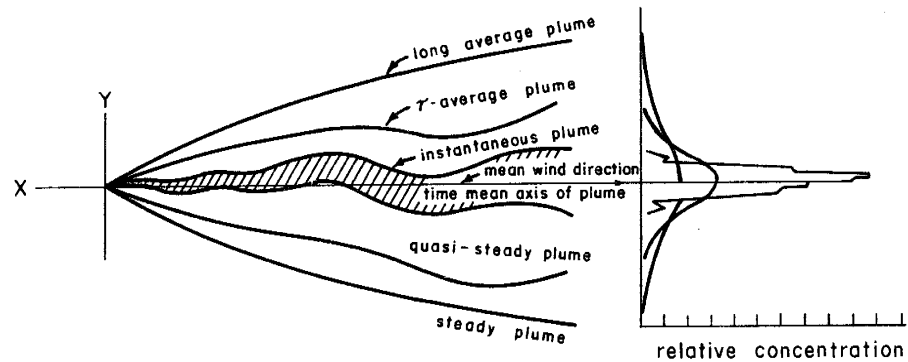


Figure 4.6 Plumes and their concentrations. The left-hand side represents instantaneous plume, quasi-steady plume, and steady plume. The right-hand side shows the corresponding cross-plume distribution of concentrations.

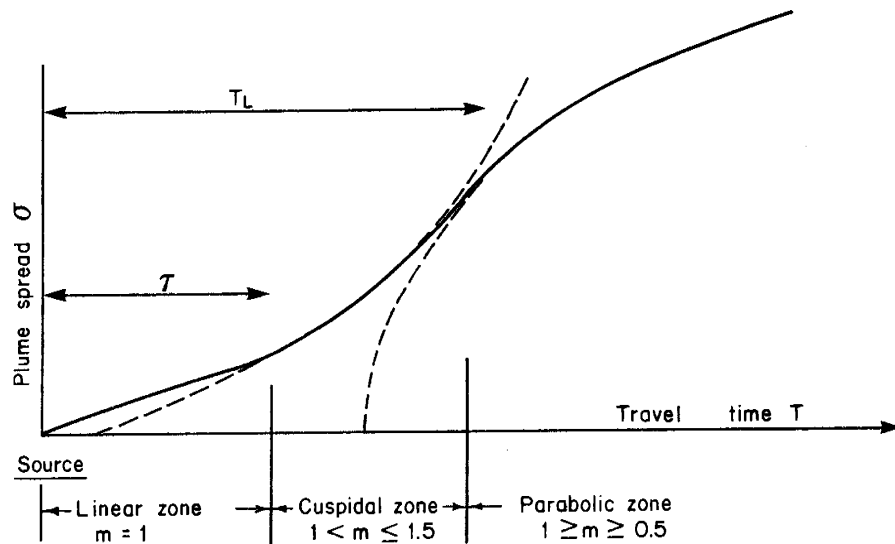


Figure 4.7 Schematic representation of quasi-steady plume. The linear zone extends into the cuspidal zone with an increase of the observation interval.

Table 4.11 Classification of turbulent diffusion from a fixed point source (Ide 1971).

Source Type	Continuous Source	Instantaneous Source
Duration of emission T_e	Large ($T_e \gg T + \tau$)	Small ($T_e \ll \tau$)
Observation time τ	Behavior Diffusion zone (T) Small $T < T_L$ Moderate $T < T_L$ Large $T \geq T_L$	Behavior Diffusion zone (T)
Small ($\tau < T$)	Instantaneous plume Cuspidal zone $1.5 \leq m < 1$	Puff Parabolic zone $1 \geq m \geq 0.5$
Moderate ($\tau \approx T$)	Quasi-steady plume $m=1$	Overlapped puff Parabolic zone $1 \geq m \geq 0.5$
Large ($\tau \gg T$)	Steady plume Linear zone $m = 1$	Overlapped puff Parabolic zone $1 \geq m \geq 0.5$

Notation:

$$\sigma \propto T^m$$

 σ^2 is variance of particles

T is travel time

 T_L is Lagrangian integral time scale

ensemble mean relative diffusion is obtained. If the average concentrations at fixed points on the ground are determined from traverse profiles, then an estimate of the total diffusion is obtained, with an averaging time equivalent to the interval between first and last traverse. Similarly, a snapshot of a plume gives instantaneous or relative spreads while the average of several snapshots will give a time average or total spread. The total spread can be decomposed by adding the variances of the relative and flapping (or displacement) component:

$$\sigma^2 = \sigma_r^2 + \sigma_d^2$$

In tracer studies in the state of Washington (Nickola 1976) the ratio between variance due to relative diffusion and total variance ranged from less than 0.10 to more than 0.80 at a distance of 800 m from the source, depending on averaging time but not on stability. In tracer studies near Calgary the ERCB (1990) reported ratios ranging from 0.04 to 1.0 for 0.5-h periods and 0.07 to 0.91 for a 1-h period. Near ratios were 0.25 and 0.22 respectively.

Other data (Plate 1982) show that single particle diffusion dominates the total variance for travel times up to about 200 s while relative diffusion is dominant beyond 1000 s. Rowe (1981) predicted that for a 1-h averaging time, relative diffusion becomes dominant after a travel time of 1500 s. A transition to the parabolic region at a travel time of 10 - 15 h has been reported (Randerson 1972); however, in the cuspidal zone growth was not as rapid as the 1.5 power expected from theory. Historical data (Gifford 1977) suggested that the accelerated region existed out to a travel time of 1000-3000 s. The accelerated growth phase of Australian plumes extended from about 2 h to 15 or 20 h (Gifford 1977), but there were too few data points to define the transition point (Barr and Gifford 1987). The existence of any extensive region of acceleration has been questioned because little evidence could be found and to support exponents as large as 1.5 (Pasquill and Smith 1983). This large value is expected only for inertial subrange turbulence which has properties unlike those of atmospheric turbulence in general. The interpretation of experimental data is further complicated by the fact that shear-induced mixing leads to a similar accelerated growth.

4.5.2 Peak-to-Mean Ratios

The difference between relative and time-averaged diffusion led workers to study the ratio of "peak" (instantaneous plume) to "average" (Eulerian fixed-point

time-average) concentrations. A differential equation describes the variation of sigma with sampling time (Singer et al. 1963), which leads to the equation

$$P/M = 1/\tau^p$$

where P is the instantaneous value of the concentration, which in practice is smoothed over a time interval usually determined by the response of the measuring instrument, M is the mean concentration over any specified averaging interval τ , p is an empirical constant.

This was later generalized to the relationship

$$C_1/C_2 = (\tau_1 / \tau_2)^p$$

where C_1 corresponds to the concentration averaged over time τ_1 and C_2 corresponds to the concentration averaged over time τ_2 .

A fluctuating plume model (Gifford 1959) provided the following conclusions:

- (a) On the plume axis peak-to-mean ratios may be 1 to 5.
- (b) With sources 50 - 100 m high, peak-to-mean ratios near the ground may be as large as 50 - 100.
- (c) With increasing distance downwind from an elevated source, ground-level peak-to-mean ratios decrease to the lower limit of unity at large distance.
- (d) Peak-to-mean ratios for off-centreline positions are greater than for centreline locations, but the difference decreases with increasing downwind distance.

These calculations yield a p of 0.2 for centreline locations and a p of 0.75 for off-centreline locations (Strom 1976).

Tracer diffusion data lead to the conclusions (Ramsdell and Hinds 1971) that:

- (a) The distribution of short-term concentration is highly skewed, with a large percentage of concentrations less than the local mean at all cross-wind plume positions.
- (b) The intensity of fluctuation of concentrations normalized to the local mean is a minimum on the mean plume axis and increases as a function of distance (the absolute intensity of fluctuations decreases, however).
- (c) Peak-to-mean ratios are a function of both cross-wind position in the plume and the time ratio, τ_1/τ_2 .
- (d) The peak-to-mean ratio monotonically increases with time ratio, but the relationship is not adequately described by a single power law.
- (e) Peak-to-mean concentration ratios at the plume centreline were higher than those shown by Gifford.

Both stability and terrain roughness affect the peak-to-mean ratios (Singer et al. 1963). Forested areas reduce the peak-to-mean ratios for sampling times less than 3 minutes. This can be attributed to the more effective mixing and reduced mean wind speed over the rough wooded surface. Values of p were found to be about 0.6 for very unstable, 0.5 for unstable and 0.3 for neutral conditions. Over rough terrain the value of p was about 0.15. Urban area data from Meade (English) and Wipperman (German) out to sampling times of 4 days followed a p value of about 0.15. The short-term fluctuations were smoothed out and obliterated because of the roughness of the city.

Theoretical and experimental work in Japan (Hino 1968) found that $p = 0.5$ was valid for sampling times from 10 minutes to 5 h. For times less than 10 minutes a value of $p = 0.2$ seemed to be valid. Atmospheric stability had little effect except on the range of applicability, the p -value of 0.5 extending to some 5 h in unstable conditions, but only to 3 h in neutral conditions. Three sets of industrial data in the U.S. yielded $p = 0.5 \pm 0.2$ for averaging times out to one year (Herman 1980). Confusion arises because various authors use different definitions of peak and mean.

TVA data showed that the ratio of peak 5-minute SO_2 concentrations to 1-h averages varied between 1.07 and 12.1 and peak 5-minute to 24-h averages varied between 1.78 and 22.3 (Montgomery et al. 1971). The ratios decreased with distance and atmospheric instability. The best estimate of the mean ratios were 2.28 for peak to 1-h, and 24.3 for peak to 24-h. The ratio was found to be a linear function of averaging time out to 3 h (Montgomery and Coleman 1975). TVA data have been interpreted as a progressive increase in the value of p with increased sampling time: for the correction of 1-minute samples to 1 h the implied value of p was 0.17; for the correction from 1 h to 3 h the implied p value was 0.37 and from 1 to 24 h it was 0.58 (Strom 1976).

Leung (1972) reported that the ratios of 3-minute to 1-h concentrations found in the literature of the 1960s gave values between 1.5 and 10 with the majority lying between 2 and 3. Using measurements made in the vicinity of the Balzac gas plant northeast of Calgary, Leung found ratios between 1.5 and 2.5 with a mean value of 2.0. This result is consistent with a p value of 0.2. The ERCB (1990) reported peak-to-30-minute ratios of 3 to 7 and peak-to-3-minute ratios of 1 to 3 from stationary tracer measurements.

It is often assumed that the averaging time correction arises only from the flapping of the plume laterally. Vertical movements are limited because the presence of the earth's surface restricts the maximum eddy size to the height above ground. This leads to the relation

$$\sigma_{y_z}/\sigma_{y_1} = \sigma_{\theta_z}/\sigma_{\theta_1} = (\tau_z/\tau_1)^q$$

Experimental studies suggest $q = 0.25$ to 0.3 is a reasonable approximation for $1 \text{ h} < t < 100 \text{ h}$ (Csanady 1973, Gifford 1975). From 1 h to a few minutes most authors agree on a value of 0.2 as originally suggested in the 1950s. For elevated sources out to a few kilometres (Gifford 1975) or sampling times up to about 10 minutes (Pasquill 1976) a similar increase with time can be expected for the vertical sigma. Thus the value of $p = 0.5$ found by Hino reflects both the σ_y and σ_z variation with time and should not be expected to extend to any greater downwind distances. A value of $q = 0.25$ has been found at distances between 10 and 68 km downwind with a slightly larger value in unstable conditions (Mueller and Reisinger 1986). The ERCB (1990), using 279 h of sonic anemometer measurements near Calgary, found that $q = 0.19$ for averaging times out to 2 minutes and $q = 0.28$ for averaging times from 3 minutes to 60 minutes.

Simple power law relations will be a satisfactory approximation only over some limited range, depending on the precise form of the spectrum. For one common spectrum, $q = 0.2$ will apply in the range $20 < u\tau < 400$ m where u is the wind speed and τ is the averaging time (Pasquill 1974). Outside this range and for the more irregular individual spectral shapes that are often encountered, especially at low frequencies, this simple law cannot be expected to apply. Furthermore, when such relations are used to estimate concentration ratios, the result must depend to an important degree on the relative positions of source and receptor.

4.5.3 Frequency Distribution of Concentration

The peak-to-mean ratio has been criticized as an inefficient statistic because it uses only one of perhaps thousands of values (Barry 1969, 1971, 1977) thus yielding a crude empirical estimate of the concentration which is exceeded with frequency $1/N$ where N is the sample size. In general the ratio P/M increases as the number of samples increases. Therefore, it is much better to use all of the information to define the distribution function which can then be used in a predictive sense for the population of possible values (e.g., Högstrom 1968). Chalk River data show (Barry 1977) that concentrations measured at a fixed point relative to a single source with averaging times of less than 100 h follow an exponential distribution

$$F(C) = F(o) \exp (-F(o)C/M)$$

where $F(C)$ is the frequency of exceeding concentration C , M is the mean concentration and $F(o)$ is the frequency of non-zero concentrations.

For averaging times longer than several hundred hours the distribution will be normal, and for intermediate times a log-normal distribution may be expected (Barry 1977). Simple theoretical arguments lead to either a normal or a log-normal distribution depending upon sampling time and decay time of the autocorrelation coefficient. Gifford's fluctuating plume model (Gifford 1974a) predicted that the logarithm of the concentration (rather than the concentration itself) is distributed exponentially which is the same as Chi-square with two degrees of freedom. The abundant literature on log-normal distributions and gamma distributions for urban areas/area sources (Berger et al. 1982) is not relevant to the discussion of individual plumes from point sources.

Following simple theoretical arguments a log-normal distribution may be obtained (Csanady (1973))

$$F(C/M) = 1 - 0.5 \left(1 + \operatorname{erf} \frac{\ln C/M}{\sqrt{2} \sigma_1} \right)$$

where F is the frequency of exceeding concentration C , σ_1 is the logarithmic standard deviation and erf is the error function. Wilson (1982) reported other theoretical and empirical support for the log-normal distribution including wind tunnel data taken at the University of Alberta by Netterville (1979). Typical values for σ_1 range from 0.2 for smooth terrain to 0.7 in rough terrain (Csanady 1973).

4.6 BUOYANCY

The growth of the cross section of a hot effluent plume over a considerable part of its rise is caused by the entrainment of ambient air resulting from the relative vertical motion of the plume. Theoretical, laboratory and observational evidence points to a linear growth of plume radius with height. This means that the vertical dimension of the plume is proportional to the plume rise, in contrast to the spread of a passive plume whose growth is dominated by atmospheric turbulence (Pasquill 1975). This initial growth may be especially important in stable conditions. However, the maximum ground-level concentration *per se* may not be greatly affected since the spread to ground level depends predominantly on the passive (atmospheric) part of the process. The condition for maximum ground-level concentration is roughly $\sigma_z = (h_s + \Delta h)/\sqrt{2}$ where h_s is the stack height and Δh is the plume rise. The induced σ_z is about $\Delta h/3.5$ which obviously cannot be the major contribution to the required sigma even when the plume rise is large.

The foregoing conclusion applies, of course, to flat terrain. In hilly topography the maximum may occur on elevated terrain much closer to the source and the plume dimensions at this distance may well be dominated by buoyancy effects (Egan and Bass 1976).

The vertical spread estimate can be corrected by means of the formula (Pasquill 1976)

$$\sigma_z = (\sigma_z^2 + \Delta h^2/10)^{1/2}$$

Buoyancy-induced spread can also be accounted for satisfactorily by the introduction of a virtual source upwind of the physical source (Moore 1974). In a qualitative sense, this is equivalent to changing the exponent in the power law for vertical spread in the vicinity of the maximum ground-level concentration.

There is also a contribution to spread from the variations in height of rise associated with relatively low-frequency variations in wind speed. The root mean square variation in h_r is approximately σ_u/u , where σ_u is the standard deviation of the low frequency (averaged over about 30 s) wind speed variations (Pasquill 1975). Since σ_u/u can exceed 0.1, the extreme range of h_r is likely to be a factor of 5, an important contribution to time mean spread. Such a contribution could add as much as 50 to 100% to the estimate of overall vertical spread (Egan and Bass 1976). Fanaki et al. (1979c) reported that in all of the studies of plumes from oil sands sources enhanced spreads occurred close to the stack as a result of buoyancy.

Wilson (1979b) expressed the buoyancy effect in the form $\sigma_z^2(\text{total}) = \sigma_z^2(\text{passive}) + \beta^2/4 \Delta h^2$ where β is the entrainment constant used in plume rise theory and the initial plume radius at the source is assumed negligible.

Davison and Leavitt (1979) concluded that $\beta = 0.6$ described the Suncor summer data quite well. They noted, however, that a power law $\sigma_z = 0.32 \Delta h^{2/3}$ actually provides the best fit to the data, and that this reduction might be the result of local updrafts.

During the rise phase, a plume cross section is assumed to be circular, which means that the horizontal spread will be enhanced by a similar amount and can be described similarly. This increase will always act to reduce the ground-level concentration. However, Davison and Leavitt (1979) reported that Lidar observation at Suncor gave horizontal spreads 1.4 times larger than vertical.

4.7 PLUME HEIGHT

For plumes rising to some hundreds of metres the prospect of systematic changes with height in the natural diffusive action becomes more likely, as does the prospect of encountering more or less sharp discontinuities in the flow structure. The latter may be responsible for some of the wide scatter in experimental data which confuses attempts to understand the systematic changes (Pasquill and Smith 1971). This has been mentioned by all those studying industrial plumes in Alberta.

The fact that the growth functions for vertical and horizontal spread are different in exponent also has some significant ramifications. If the exponents are the same, as is commonly assumed for simplicity (or as is contained in the BNL curves), the maximum ground-level concentration is proportional to the reciprocal of the square of the effective stack height. However, with different growth rates the maximum ground-level concentration is proportional to the reciprocal of $h_{eff}^2 (1 + d/b)$. Since d is usually larger than b an increase in source height is even more beneficial than is commonly assumed (Pasquill 1968, 1971).

The height dependence of the various sigma curves has been identified by many workers (Draxler 1976, Briggs 1973, Ragland 1976). Height-dependent sigma curves have been developed from extensive tracer experiments in the Federal Republic of Germany (Vogt 1977): in neutral and slightly unstable conditions, the sigmas decreased with height; in moderate to strong instability the sigmas increased with height. Markov chain simulations show that vertical sigma is fairly independent of height in the neutral surface layer (Reid 1979). In unstable daytime conditions, the horizontal sigma varies little with height, but the vertical sigma increases by a factor of 2 to the mid-height of the planetary boundary layer (Hanna 1981). The vertical sigma has been found to increase with height in neutral and stable conditions for short downwind distances (Doran et al. 1978). In unstable conditions the lateral spread has been found to be independent of height and the vertical spread weakly dependent (Lamb 1979). Leahey and Hansen (1985) found that the ratio of vertical to horizontal turbulence decreased rapidly with height in all stabilities.

Experimentally determined vertical spread values for plumes about 250 m from tall stacks in Japan were all between the C and D curves in the Pasquill-Gifford

system (Senshu et al. 1974). The representative stability corresponding to the experiments was near neutral irrespective of the stability of the surface boundary layer. Similar results have been reported over flat terrain (Sakuraba et al. 1969), but over complicated terrain the elevated source shows larger spreads. When released below an elevated stable layer the plume was observed to split and spreading in the lower layer was larger than in the upper (Sakuraba et al. 1973, 1972).

Kumar (1979b) used a gradient transfer model with semi-empirical eddy diffusivity profiles to develop sigmas for arbitrary release heights. Values of the vertical spread decreased as the release height increased from 100 to 500 m under all stabilities. Release heights lower than 100 m also showed a reduction except in the stable case owing to the lower level of turbulence. In contrast, an analytical solution of the gradient transfer equation and a least-squares fit to calculate the effective Gaussian sigma and wind speed, showed that the vertical sigma increased with source height, while u decreased (Melli and Runca 1979).

Yet another difficulty arises in the choice of mean wind for use in the Gaussian plume model. The normal 10-m "surface" wind is recommended (Pasquill 1962) for distances of 0.1 and 1 km. For distances of 10 and 100 km, a speed midway between this and the geostrophic value estimated from a chart of the surface pressure distribution should be adopted. Sometimes it is recommended that the mean wind speed through the vertical extent of the plume should be used (Turner 1970), that is, the average wind speed from $h_{eff} - 2\sigma_z$ to $h_{eff} + 2\sigma_z$. If $2\sigma_z$ is greater than h_{eff} then the wind should be averaged from the ground to $h_{eff} + 2\sigma_z$.

The optimum height for estimating the mean wind speed can be found by weighting a power law wind profile with a Gaussian vertical concentration distribution (Smith and Singer 1966). This "mean equivalent wind" was found to be that estimated by a power law at a height of about $0.62\sigma_z$. The wind at source height should be used until $0.62\sigma_z > h_{eff}$. The increase of wind with height tends to decrease ground-level concentrations. The use of mean equivalent wind reduces the values most significantly for sources near ground level.

A Monte Carlo diffusion model (Reid 1981) gave the result that in neutral

conditions the mean wind should be specified at a height of $0.6\sigma_z$ or $0.6 h_{\text{eff}}$, whichever is larger, to enable σ_z to be independent of height up to plume heights of 150 m and downwind distances of 10 km.

Djurfors and Netterville (1978) showed that the height at which the wind speed ought to be evaluated to obtain the correct plume rise is a complicated function of atmospheric and stack parameters. In practice the average between the wind at stack height and plume height is often used. Wilson (1979b) avoided integrating the wind profile by using the wind speed at plume height, but found that this still underestimated the plume rise. Wilson (1981) proposed that the effective height be taken as $H + 0.17 \sigma_z$ for elevated sources. For ground-level sources he showed that the wind should be evaluated at $0.165\sigma_z$.

The need for a specific reference height for u can be removed (Pasquill 1976). The denominator of the Gaussian plume equation contains the product of σ_y and u , both of which must be evaluated at some effective height. At short range $\sigma_y = \sigma_v T = \sigma_v x/u$ which cancels u from the equation. σ_v does not change markedly with height except in stable conditions. It has been reported (Doran et al. 1978) that σ_θ measured near ground level is more useful in describing tracer results than a value measured at plume height. The same tracer results showed that the vertical spread was an increasing function of height in neutral and stable conditions.

As a plume spreads vertically and continues to grow it eventually encounters a "lid", or elevated stable layer which more or less terminates the upward growth. The vertical spread values must be adjusted accordingly. Vertical growth may be stopped when $\sigma_z = h_{\text{eff}}/2.15$ (Meade 1960), but this assumes a cessation of mixing while a considerable vertical concentration gradient remains (Smith and Singer 1966). If it is assumed that the vertical distribution of material within the layer should be nearly uniform with height, then σ_z should become a constant when it is equal to the depth of the mixing layer (Pasquill 1962). The multiple reflection formulation of the Gaussian plume model obviates the need for such corrections in models. However, for experimental sigma determinations Whaley (1974) and Rowe (1975) have pointed to the importance of the relationship.

Because plume material will experience strongly varying turbulence as it spreads under stable conditions, Wilson (1987) used weighted values over the plume layer

to generate effective sigmas that account for the fact that individual plume elements spend a larger fraction of time near the plume centreline. This was incorporated in the sour gas concentration model of Alp et al. (1990).

4.8 LONGER TRAVEL TIMES

Most atmospheric diffusion studies have been concerned with travel times of less than 4 h. Early data on horizontal plume spread at longer distances from a wide variety of sources yield $\sigma_y(\text{km}) = 1.8T(\text{h})$ (Heffter 1965, Gifford 1983). Plume spreads out to 250 km in neutral stability have been reported to follow an extension of the Pasquill D curve (Peterson 1968), with a best-fit equation of $\sigma_y = 0.48 x^{0.96}$. Plume spreads at distances up to 560 km under cloudless, convective conditions over flat terrain gave the relation $\sigma_y = 0.334 x^{0.884}$ (Bigg et al. 1978) which is similar to the Pasquill curves, and to the $x^{0.8}$ dependence derived from reexamining older long-distance diffusion data (Pasquill 1974).

A replot of earlier data on relative diffusion (Gifford 1977) led to the conclusion that overall the best fit gave the horizontal spread proportional to $T^{1.2}$. Adding other data gave evidence of two separate regimes. Out to around 1000 - 2000 s the spread varies at $T^{3/2}$, but slows thereafter in the region of the mesoscale spectral gap.

A plume or puff first experiences boundary layer turbulence, its growth controlled by larger and larger eddies as time (and travel distance) increases. After several hours it grows larger than the depth of the boundary layer and after several days may spread through most of the troposphere. The character of the motions causing the diffusion changes markedly as boundary layer turbulence gives way to the two-dimensional "turbulence" of large-scale systems. Relative spread varying as $T^{3/2}$ is expected in the inertial subrange, as opposed to the average spread which varies as T . At still larger times a second region of accelerated growth is anticipated when the energy cascade from synoptic disturbances is encountered (Gifford 1977).

Data from Australian can be represented by the function $\sigma_y = 1.4T^{0.995}$ which is very similar to Heffter's results (Gifford 1983). Three regions can be identified: (1) a roughly linear region from 0.1 - 2 h; (2) a region of accelerating diffusion from 2 - 20 h; and (3) a region of decreasing diffusion. The first region is the result of small-scale

turbulence acting on the edges of the plume; larger-scale PBL turbulence acts only to displace the plume (affecting time-averaged but not relative spread). The second region is associated with PBL and larger-scale processes.

Few data approach the square root growth predicted by Taylor's theory for long travel times (Randersen 1984). This may be attributable to wind shear in the boundary layer. Models that include shear-induced spread predict growth rates similar to $T^{1.2}$.

A theory with three parameters independently related to the source configuration and the turbulence structure has been developed (Gifford 1983):

$$\sigma_y^2 = \sigma_{y0}^2 + 2 v^2 t_L^2 [T/t_L - (1 - e^{-T/t_L}) - 0.5c(1 - e^{-T/t_L})^2]$$

where v^2 is the mean turbulent wind velocity, t_L is the Lagrangian time scale, T = travel time, σ_{y0} is initial source size and c is a parameter that expresses the effect of source configuration on initial space-and-time wind correlations. The parameter c affects the early travel times whereas longer times are dominated by t_L and v^2 . For short-duration sources of small extent, $c = 1$ and there is an initial accelerated growth phase followed by a $T^{1/2}$ regime. Long-time average plumes from continuous sources are characterized by $c = 0$ which leads to an initially linear growth, followed by a $T^{1/2}$ regime. Intermediate values of c are associated with sources extended in time or space. Thus, the equation is consistent with earlier theoretical results. This equation fits the Australian data better than T , $T^{1/2}$ or $T^{0.91}$ power laws (Gifford 1982). A value of $t_L = 10^4$ and v of 1-2 m/s has been derived from 12 data sets (Barr and Gifford 1987).

To some extent this resolves the criticism (Pasquill and Smith 1983) that the existing data did not give a clear indication of accelerated growth as large as $T^{3/2}$, although it was also argued that the regime would be of limited extent because it applies only in the inertial subrange. A variation proportional to $T^{0.8}$ (Pasquill 1974) was later revised with newer data to 0.875 (Pasquill and Smith 1983). The maintenance of this rate and the lack of transition to a $T^{1/2}$ regime was attributed to the effect of wind shear in the boundary layer.

Few observations of longer range plume spread have been made in Alberta or even in Canada. Millan and Chung (1977) detected the Sudbury plume 400 km downwind

using a correlation spectrometer. The spreads were consistent with values predicted by extrapolating the ASME (BML) curves for the inferred stability. Cheng et al. (1987) measured the spread of the Syncrude plume at distances out to 70 km, and reported downwind growth as $x^{0.8}$.

4.9 WIND SHEAR

The variation with height of the mean wind direction and speed displaces the plume centreline, distorts the cross-sectional shape and enhances the plume spread.

The most obvious effect is that the plume centreline projected on a horizontal surface appears curved. This alters the location of the maximum ground-level concentration and can cause problems with photographic determinations of plume parameters. Davison and Leavitt (1979) showed that the displacement can be approximated by

$$Y_c = \frac{X}{2} \tan \alpha$$

where X is the downwind distance to final plume rise and α is the change in wind direction over the layer from stack top to plume top.

Because of the change in wind direction with height, the upper and lower portions of the plume move in somewhat different directions. This causes the elliptical plume cross section to become tilted. The tilt itself is of little concern in dispersion modelling but it is a source of difficulty in field measurements of plume spread. Hoff and Froude (1979) and Slawson et al. (1978) showed numerous plots of tilted, elliptical plumes in the oil sands area.

The tilt leads to enhanced spread when vertical mixing occurs, because material moves upward from the lower edge and downward from the elevated edge. Davison and Leavitt (1979) estimated the contribution to the spread as

$$\sigma_s = 0.29 \sigma_z x \frac{\tan \alpha}{\Delta Z}$$

where α is the turning angle of the wind over depth ΔZ and x is the downwind distance.

Aircraft measurements of the Suncor plume showed that wind shear occasionally made a significant contribution to plume spread, but only at distances beyond 10 km. This is supported by the theoretical results of Kumar (1979b).

Shear-enhanced diffusion is likely to be important only in moderate stability; strong mixing will eliminate the shear and strong shear is associated with weak mixing. The maintenance of plume spreading at the rate of $x^{0.875}$ rather than the theoretical limit of $x^{0.5}$ could be due, at least, in part, to the wind directional shear across the boundary layer (Pasquill and Smith 1983).

For distances beyond 20 km, a rough rule based on limited observational data calls for the addition of a component of horizontal variance given by $0.03 \Delta\theta^2 x^2$, where $\Delta\theta$ is the turning of the wind direction in radians over the entire plume (Pasquill 1976).

4.10 SURFACE ROUGHNESS

Since mechanical turbulence is controlled by the roughness of the underlying surface, it follows that plume spreads will also be affected by surface roughness. This was first manifest in the diffusion rates observed for tracer experiments in cities (McElroy and Pooler 1968). A set of revised plume spread curves based on experiments over urban, suburban and wooded areas has been proposed (Bowne 1974). These curves effectively shifted the Pasquill curves by 1 - 2 stability categories to the unstable.

The revised Pasquill system provides separate vertical spread curves for three different orders of magnitude of roughness length (Smith 1972). The results of diffusion experiments conducted at two sites in Germany (Nester et al. 1977) were broadly in agreement, although the measured values tended to be larger (Miller 1978).

Kumar (1979b) reported that the effect of surface roughness in his numerical modelling results could be summarized by

$$\sigma_{y_2}/\sigma_{y_1} = (z_{02}/z_{01})^r \text{ where } r \text{ is } 0.05 \text{ to } 0.2$$

$$\sigma_{z_2}/\sigma_{z_1} = (z_{02}/z_{01})^s \text{ where } s \text{ is } 0.05 \text{ to } 0.1$$

A similar relationship with an exponent of 0.2 has been reported in the literature (Hanna et al. 1977).

An increase in the vertical sigma brings an elevated plume to the ground more quickly, causing higher concentrations. The maximum ground-level concentration shifts toward the source, and concentrations at larger distances are less than with a smoother surface. For ground-level sources, larger roughness always decreases concentrations everywhere.

Buildings and other large obstacles also act to increase the spread rates. These increases can be formulated into sigma enhancements using the characteristic length scales of the building (Hosker 1984).

Although the effect of a change in surface roughness on airflow characteristics has been studied extensively, for example, Angle (1975), the effects on plume dispersion have seldom been investigated. Wind tunnel results have demonstrated that in neutral conditions, ground-level concentrations are not much affected because of the compensating effects of reduced wind speed and larger turbulence (Pendergrass and Arya 1984a,b). Larger-scale upstream turbulence dominates any effect from roughness change. Further discussion of surface roughness effects can be found in Chapter 5.

4.11 ALONG-WIND SPREAD

The Gaussian plume equation assumes that the turbulent spread of material in the direction of the wind is negligible compared to the transport of material by the mean wind. In any event, diffusion in the along-wind direction will simply redistribute mass to different downwind locations with no net change in the mass flux over a given location. However, if the source strength varies in time, or if the source is of short duration, or if the wind speed is very low, along-wind diffusion becomes important (Ukeguchi et al. 1973).

A common method of dealing with the three situations above is to employ the Gaussian puff equation (Gifford 1968)

$$C = \frac{2Q}{(2\pi)^{3/2} \sigma_y \sigma_z \sigma_x} \exp [-0.5(y/\sigma_y)^2] \exp [-0.5(z/\sigma_z)^2] \exp [-0.5 (x - uT)^2/\sigma_x^2]$$

where Q is the total quantity (mass or volume) of material released. At the centre of the puff $x = uT$ and the last exponential becomes 1.

In most applications of the puff model, the along-wind spread is assumed equal to the cross-wind spread (e.g., Ludwig et al. 1977). Field observations (Nickola 1971) have shown that irrespective of stability, the along-wind spread is larger than the horizontal spread by factors ranging from 1.3 in unstable conditions to 4.3 in stable conditions. It has been concluded that the along-wind spread is about a factor of 2 larger than cross-wind spread at longer travel times (Draxler 1979), which may be attributed to both the along-wind turbulence intensity and wind shear.

Wilson (1979b, 1981) expressed the along-wind spread as

$$\sigma_x = [\sigma_{x \text{ shear}}^2 + \sigma_{x \text{ turbulence}}^2]^{1/2}$$

and expressed the shear term as

$$\sigma_{x \text{ shear}} = 0.3 (n x/z_r)(z_r/z_c)^n \sigma_z$$

where z_r is the reference height for wind, n is the index of the power law for the wind profile and z_c is the convection height.

4.12 NONSTATIONARITY AND CALMS

When a source is of short duration or changes its strength over time or when mean wind and turbulence levels are changing, the straight-line conical plume concept is no longer valid. Two basic approaches have been developed:

- (1) segmented plume models (Hales et al. 1977, Benkley and Bass 1979, Lott 1977) divide the plume into a number of independent sections, each of which has different characteristics. The concentration at any point is calculated using the plume equation for a virtual plume that starts at a virtual source and passes through the closest plume segment.
- (2) multiple puff models (Roberts et al. 1970 a, b, c; Lamb 1968; Shieh 1969; Shieh et al. 1970) represent the emission as a series of instantaneous point sources and compute the concentrations by superimposing the contribution of each puff.

Puff models are extremely versatile in that they can handle not only changes in time, but also changes in space and low wind speeds. Many features have been incorporated including: (1) wind shear effects (Shieh 1978); (2) virtual distance (Ludwig et al. 1977) and virtual age (Zannetti 1981) for the correct evaluation of sigmas; (3) puff merging and purging (Ludwig 1984, Ludwig et al. 1977) or puff splitting (Zannetti 1981) to reduce computer requirements; and (4) time-dependent sigmas (Roberts et al. 1970 a, b, c; Zannetti 1981, 1986). Choukalos (1978) implemented an integrated puff model for early studies of sour gas blowouts as did the ERCB (1979). Wilson (1979 b, 1981) described an effective plume model for situations involving an exponential transient source such as a pipeline blowout. This was incorporated into a risk model by Choukalos (1980).

One of the difficulties in applying puff models is the specification of sigmas since relative rather than total values are required. Hage et al. (1966) summarized 16 experimental determinations of horizontal spread and 5 of vertical spread from puff releases. Despite these and other studies (Slade 1965, Islitzer and Slade 1968, Draxler 1984) few models use puff sigmas (Daggupaty 1980, IAEA 1976). Most models still employ plume sigmas (Petersen 1986, Scire et al. 1984, Zannetti 1986). Puff models are often used to estimate the dosage, defined as the integral of concentration over time (Daggupaty 1980, Csanady 1969). If the puff passes rapidly overhead so that the sigmas are virtually constant throughout the time of passage then the equation for dosage or exposure has the same mathematical form as that for a continuous plume (Gifford 1968) but with relative, rather than total sigmas. Johnson (1967) reported good agreement between plume predictions and measured dosages from instantaneous phosgene releases at Suffield.

The distinction between a puff and a plume depends on travel time and release time. When the travel time exceeds the release time, the "plume" is detached from the source and thereafter travels and grows as an elongated cluster or puff (Pasquill 1974). The ratio between plume estimated concentrations and puff estimated concentrations is given by

$$C_{\text{plume}}/C_{\text{puff}} = (2\pi)^{1/2} (\sigma_x/ut_r)$$

where t_r is the release time. For short-duration releases ($t_r < 2.5 \sigma_x/u$) the plume model will predict higher values, whereas for longer releases the puff model will predict higher

values. This led Wilson (1986) to recommend that the correct prediction will be obtained by using the model which predicts the lower concentration.

Under low wind speed (calm) conditions most workers have found that time-average concentrations are lower than those predicted by Gaussian plume models because of the meander of the wind (van der Hoven 1976, Wilson et al. 1976, Sagendorf and Dickson 1976, Draxler 1984). A segmented plume approach, updating the meteorological conditions every few minutes, seems to give good results (Sagendorf and Dickson 1976) as do multiple puff models (Roberts et al. 1970 a). Gradient transfer theory (Ukeguchi et al. 1973) and puff-based analytical solutions (Yamartino 1977) have also been applied.

4.13 CONCENTRATION FLUCTUATIONS

Like other atmospheric boundary layer variables, chemical concentration exhibits turbulent fluctuations. In a typical time series of concentrations, fluctuations are at least as large as the mean. Such variability is important because lethality, toxicity and flammability are related to short-term peak concentrations. These fluctuations also impose an expected uncertainty in any comparison of field measurements to model predictions. Part of the fluctuation is due to incomplete mixing in the plume and part is due to the meandering or flapping of the plume about the measurement point. The latter occurs when the scale of the atmospheric eddies is larger than that of the plume. The mixing of a puff of material in the ambient air can be envisioned as resembling a loosely tangled ball of wool in which high-concentration filaments, referred to as the "core", are gradually distorted and pulled apart into pockets of low concentration, referred to as the "bulk" (Chatwin and Sullivan 1979). Seven different approaches for describing and estimating fluctuations are summarized in Table 4.12 (Hanna 1984b).

Some of the leading work on concentration fluctuations has been done in Alberta. Netterville (1979) measured concentration variance profiles in a wind tunnel and showed that a Gaussian distribution provided a useful approximation if the positive image source were replaced by a negative image sink term. He also demonstrated that a Gaussian profile results from a variance balance equation under special circumstances which, based on the wind tunnel data, are not unreasonable. The dissipated Gaussian model was used by Netterville and Wilson (1980) to predict concentration probability

Table 4.12 Summary of approaches to concentration fluctuations (after Hanna 1984b).

Model Type	Approach	Main Parameters
K	gradient transfer of variance	dissipation α diffusivity κ
Gaussian	conditional averaging Gaussian spatial distributions	intermittency γ conditional mean c_p conditional variance σ_p^2
Similarity	core/bulk structure	puff dimension σ source size σ_o
Meandering Plume (Gifford 1959)	total variance is sum of variances due to plume spread and meandering	disc spread σ_M plume spread σ_p
pdf (Hanna 1984a)	concentration fluctuations directly related to turbulent velocity fluctuations	averaging time T sampling time S integral time scale T_z touchdown velocity $\bar{W}$
Monte Carlo	rate of separation is a function of instantaneous separation	Eulerian length scale L source size σ_o
Advanced Numerical	Second-order closure or large eddy simulation	third-order statistics and functional relationships

distributions. The interplay between models and test data and the evolution of our understanding of concentration fluctuations is evident in the sequence of works Wilson (1979b, 1982, 1986), Wilson et al. (1982 a, b, 1985), Wilson and Simms (1985), and Wilson and Bara (1985).

The theoretical developments are based on the concept of intermittency and the use of conditional of statistics those obtained only when source material is present, and those which include periods with no source material. Intermittency, represented by the symbol γ , is defined as the fraction of time that the source material is observed. Mean concentrations and variances measured only when the plume is present are called "conditional" or "plume" values as distinct from "total" values. It follows that

$$c_p = c/\gamma$$

$$\sigma_c^2 = \frac{1-\gamma}{\gamma} c^2 + \gamma c_p^2$$

$$i = \frac{\sigma_c}{c}, \quad i_p = \frac{\sigma_p}{c_p}$$

where the subscript "p" refers to the conditional (plume) values and the unsubscripted terms refer to total values.

The total variance along the plume centreline is given by a Gaussian equation

$$\sigma_c^2 = \left(\frac{q}{2\pi u \sigma_y \sigma_z} \right)^2 \left[\exp -0.5 \left(\frac{z-h_v}{\sigma_z} \right)^2 - \alpha \exp -0.5 \left(\frac{z+h_v}{\sigma_z} \right)^2 \right]$$

where	q	is the variance source strength
	h_v	is the variance source height
	α	is the sink strength parameter, equal to about 0.6 based on wind tunnel data
	σ_y, σ_z	are the spreads of the mean plume

The effective height of the variance source is displaced above the mean plume height and is given by

$$h_v = [h_p^2 + 0.49 \sigma_z^2]^{0.5}$$

The vertical variance of total intensity can then be calculated. At the surface, mean reflection and variance dissipation reduce the intensity by a factor of three, but do not force it to zero. The variance source strength changes with downwind distance as a result of the balance between advection, production, diffusion and dissipation of variance. It can be calculated from

$$\frac{q}{Q} = \frac{0.34 \lambda}{\lambda^2 + \lambda_o}$$

where λ is a nondimensional downwind distance defined as

$$\lambda \equiv (\sigma_y \sigma_z / h_p)^{0.5} \text{ and } \lambda_o \text{ is}$$

a virtual origin that accounts for the effect of source size on the contribution meandering to the total variance.

The meandering component depends on the ratio of source size to the average integral scale of turbulence. An empirical equation for which is

$$\lambda_o = \frac{0.12}{0.055 \Lambda / d + 1}$$

where Λ is the integral scale of turbulence, which is given approximately by

$$\Lambda = 0.6 h \exp -(\sigma_z / h_p)^2$$

Although the conditional intensity is observed to increase at the plume edges, assuming it to be constant provides a useful model which yields

$$i_p = \frac{0.34 \lambda}{\lambda^2 + (.12)^2} \text{ away from the ground}$$

The intermittency is then determined from the total and conditional intensities using the equation given earlier.

These relations suggest that conditional averages can be used to estimate centreline means. In the fringes of the plume

$$C_o = \frac{1 + i_o^2}{i_o^2} C_p$$

For typical values of i_o between 0.5 and 1.0 the conditional mean in the fringes of the plume will be 1/5 to 1/2 of the plume centreline value.

4.14 CHEMICAL TRANSFORMATIONS

As a plume of material moves downwind it may react with other gases or particulates and change into different compounds, possibly in different phases. This reduces the concentrations of the original chemical, but may produce something of even greater concern. Atmospheric chemistry is exceedingly complex, and while much is known, much remains to be learned (Bottenheim and Strausz 1977, Abel et al. 1984, Seinfeld 1986, Lloyd and Lurmann 1985). Sandhu et al. (1980) concluded that oxidation of sulphur dioxide was dominated by gas-phase reactions involving photochemically created free radicals and by liquid-phase reactions in the presence of metal ions.

Rowe et al. (1977) were unable to detect any significant oxidation of sulphur dioxide within the first 4 km downwind of a sour gas plant. Fanaki et al. (1979a) observed little oxidation of the sulphur dioxide in the plume from the Suncor oil sands plant at distances up to 30 km in the winter. This result was consistent with photochemical mechanisms, but not with the expected rate by an iron-catalyzed mechanism. Lusi et al. (1978) and Fanaki et al. (1979b) observed oxidation rates in the range of 2-3%/h during summer days and less than 0.5%/h in early morning and late evening. This further supported the dominance of homogenous gas-phase reactions and gave no support to the idea of a heterogeneous (metal-catalyzed) mechanism. The reactive plume model developed by Bottenheim and Strausz (1981) was able to reproduce the observations of oil sands plumes over the range of oxidation rates from 1 to 5%/h. Bottenheim (1982) measured a summer oxidation rate of 2-6%/h in the plume from the Syncrude oil sands plant about 9 km downwind. He also found evidence of an iron-catalyzed liquid oxidation

during a cold early night flight in winter. Cheng et al. (1987) noted that the data from Lusis et al. (1978) showed some winter and early morning summer flights with large oxidation rates, which could be interpreted as catalytic oxidation. Davis et al. (1985a) and Cheng et al. (1987) reported oxidation rates in March varying from 0.02%/h to 2-8%/h and in June varying from 0.06%/h to 8.7%/h for plumes up to a few hours old (downwind distances up to 35 km).

The conversion of nitric oxide to nitrogen dioxide has been studied in Alberta largely in the plumes of compressor engine exhausts. Angle et al. (1979) observed an NO_2/NO_x ratio which increased steadily from an in-stack value of 0.03 to a value of 0.2 at a travel time of 70 s. Measurements in Texas and elsewhere in the U.S. were reviewed by Vet et al. (1982), who also developed a model for the conversion based on a 38-step reaction mechanism, simplified from the generalized scheme of Bottenheim and Strausz (1978).

Leahey et al. (1981, 1983) measured the ratios of NO_2/NO_x in the exhaust from a turbine driven compressor near Calgary. Median values of 0.37, 0.063 and 0.55 were found in mid-summer, early autumn and early winter respectively. A simple three-reaction photochemical mechanism was able to predict the autumn and winter values, but overestimated the summer value by a factor of 2. These ratios were all larger than those observed in the exhaust from reciprocating engines and no variation with travel time was found for at least 3 or 4 minutes.

Concord Scientific (1986) examined a large amount of continuous monitoring data collected around compressor stations. Ratios from 0 to 1.0 were found and no dependence on travel time was evident. The ratio was found to increase, however, with decreasing NO_x concentration, and the correlation was used in the new regulatory model described by Angle et al. (1987) and Sakiyama et al. (1988). Davis et al. (1985a) and Cheng et al. (1987) measured conversion rates in oil sands plumes ranging from 0.2%/minute to 21.9%/minute. The rate increased with time, reaching a maximum around noon and then declined. The rate was higher in summer than in winter, but this may not have been caused by increased solar radiation since a strong correlation with ambient turbulence levels was observed.

In terms of a Gaussian plume model, the transformation of one chemical to another is often represented by an exponential source decay of the form $Q = Q_0 / e^{kt}$

where Q_0 is the initial source strength, t is time and k is the rate of conversion of the first chemical to the second. However, it is possible to incorporate a full reaction scheme as has been demonstrated by Vet et al. (1982), Bottenheim and Strausz (1981) and Fung et al. (1985).

4.15 REMOVAL

As a plume of material moves downwind it may be subject to a number of physical processes, other than turbulent diffusion, that act to reduce the concentration in the air by transferring the material to the earth's surface. As soon as the lower edge of the plume touches the ground, removal can occur by impaction, absorption, electrical attraction, adsorption, gravitational settling and chemical reaction, processes collectively known as *dry deposition*.

Gravitational settling or sedimentation of large particles ($>10 \mu\text{m}$), is most often represented as a "tilted plume" where the effective height of the plume is progressively reduced with downwind distance. Since material reaching the ground is absorbed, no image term is included and the equation for deposition becomes

$$D = v_s \chi = \frac{v_s Q}{2\pi\sigma_y\sigma_z} \exp\left(-\frac{y^2}{2\sigma_y^2}\right) \exp\left(-\frac{(h_{\text{eff}} - v_s x/u)^2}{2\sigma_z^2}\right)$$

where v_s is the mean settling velocity of the particles. Stewart and Csanady (1967) obtained good results using this equation, as did Hage (1961a, b) using a gradient transfer representation, and Walker (1965) using a simple ballistic model, to describe the deposition of glass microspheres observed during experiments at Suffield.

For all of the dry deposition mechanisms, unless the surface is saturated, the rate of deposition is proportional to the concentration in the vicinity of the receiving surface. The ratio of the deposition flux (mass per unit area per unit time) to the local chemical concentration has the dimensions of velocity and thus the normalized deposition rate is known as the *deposition velocity*. The value of the deposition velocity depends upon wind speed, stability, topography, ground cover, surface wetness and other features of the receptor. Consequently, measured values range over four orders of magnitude for gases and three orders of magnitude for particles (Sehmel 1980). The data become more

orderly if an electrical analog is used. The three main resistances to deposition are in series and may be written as

$$v_d = \frac{1}{r_t} = \frac{1}{r_a + r_b + r_c}$$

where r_a is the atmospheric resistance, which can range from 0 during unstable afternoons to 10 s/cm during very stable nights; r_b is the surface boundary layer resistance, which accounts for the mass transfer through the quasi-laminar layer of air in contact with the surface; and r_c is the canopy resistance, accounting for stomatal, mesophyll and cuticular uptake, with values ranging from 0.5 to 200 s/cm.

Resistances for various surface types and mean monthly SO₂ dry deposition velocities for much of North America have been mapped (Voldner et al. 1986). Early empirical results were summarized by McMahon and Denison (1979) and Denison et al. (1979). The state of knowledge and research needs have been reviewed elsewhere (Hosker and Lindberg 1982).

Gaussian plume models may incorporate dry deposition in three ways. The source depletion method (Chamberlain 1953) accounts for the loss of airborne material by reducing source strength as a function of downwind distance. While the method correctly calculates the deposition to the surface, it also distributes the material loss instantaneously throughout the vertical extent of the plume. This leads to overprediction of total deposition and underprediction of the remaining airborne material by as much as a factor of 4 at a distance of 10 km in stable conditions with strong deposition; however, with low deposition rates and neutral to unstable conditions, errors are only about 20% (Horst 1977). Factor-of-2 accuracy can be expected out to 100 km even in stable conditions for typical deposition velocities (Corbett 1981). Benjamin (1975), Western Research and Development (1978) and Leahey and Burns (1979) used this method to estimate annual sulphur deposition in the vicinity of sour gas plants in Alberta.

The surface depletion method (Horst 1977) uses material sinks or "negative" plumes emitted from the surface to account for deposition and expresses the local concentration as the sum of the nondepositing concentration from the primary source plus the negative contribution from all upwind sinks. Under stable conditions, the surface depletion method predicts a deposition several times smaller than the source

depletion method. The error can be corrected in the source depletion method by using an adjusted deposition velocity

$$\frac{1}{v_{dc}} = \frac{1}{v_d} + \frac{1}{u_a^{1/2}}$$

where the second term represents an atmospheric resistance correction and the denominator of this term represents an effective turbulent velocity (Yamartino 1985).

The partial reflection method (Overcamp 1976) multiplies the reflection term by a factor α (α) to account for the loss of material at the surface. Out to about 10 km the partial reflection method gives values very similar to those of the surface depletion method (Horst 1978, Overcamp 1978). The partial reflection method does not require the computation of an integral and hence is the simplest of the three methods. At large distances, however, it appears to overestimate the reduction in concentration.

In northeastern Alberta, an SO_2 deposition velocity to snow of 0.25 cm/s was measured by Barrie and Walmsley (1978). Chadder et al. (1984) measured deposition velocities to ripe barley near Calgary ranging from 0.1 cm/s in the evening to more than 1.6 cm/s in daylight hours. Legge (1988) reported a mean morning deposition velocity in late winter of 0.57 cm/s at a site northwest of Calgary, and a mean afternoon deposition velocity of 1.2 cm/s.

Clouds may act to remove plume material in two ways: *rainout* occurs when the material is entrained into the cloud droplets, either by nucleation or by scavenging, and then falls to the ground with precipitation; *washout* occurs as the precipitation falls through the plume. While dissolved in cloud droplets, the material is subject to a variety of aqueous-phase reactions which cause further transformations.

For particles and insoluble gases that are captured by water droplets, concentration (C) will fall exponentially with time(t), that is,

$$C(t) = C(o) \exp (- \lambda t)$$

where λ is called the scavenging coefficient which can range in value from $.4 \times 10^{-5}/s$

to $3 \times 10^{-3}/s$. An alternative formulation assumes an equilibrium partitioning between gaseous and liquid phases, expressed as

$$W \equiv \frac{K}{C}$$

where K is the concentration in water, C is the concentration in air and W is called the washout ratio (although it includes the effect of rainout too). Wet deposition is then given by:

$$D_w = CWR$$

where R is the rainfall rate, typically 1×10^{-6} m/s, and the washout ratio ranges from 0.1×10^6 to 1.1×10^6 for particulates and 0.01×10^6 to 0.4×10^6 for reactive gases. Analogous to dry deposition, a wet deposition velocity can be defined as

$$v_w = WR$$

which yields values of several hundreds of centimetres per second. For soluble gases, equilibrium is generally achieved between the two phases after a water drop has moved only a few tens of metres, and then

$$W = \frac{1}{H}$$

where H is Henry's law constant, the reciprocal of the solubility coefficient. For reactive gases such as sulphur dioxide, an enhanced solubility coefficient must be used, which leads to larger washout ratios (in the order of 10^4 for SO_2). The short-range deposition models of Western Research (1978), Leahey and Burns (1979) and Alberta Environment (1988) all use this equilibrium scavenging concept.

4.16 SUMMARY

The Gaussian diffusion formula for plume spread is a practical model and a convenient framework for expressing the effects of various complicating factors. Plume spreads, represented by the standard deviation of the Gaussian distribution, can be

specified by any of four major empirical systems, seven systems based on statistical theory or two systems linking other theoretical approaches. Studies of industrial plume spreads in Alberta have often found much larger values than indicated by the prediction schemes and sometimes different growth rates. This has generally been attributed to the effects of buoyancy, topography, multiple sources, wind shear, elevated inversions or stability category selection. Of the systems that have been tested in Alberta there is a suggestion that the Pasquill method (empirical) and the Draxler method (statistical theory) are the best overall performers.

Some of the discrepancies can be explained by the differences between the spread of the plume about its own centreline (relative diffusion) and the spread of the plume as observed at a fixed location on the ground, which includes movements of the entire plume (flapping). Alberta measurements suggest that total spreads can be four times larger than relative spreads. Alberta data also support the 0.2 power law for averaging time adjustments.

Buoyancy makes an important contribution to plume spread near the source. Plume spreads may be height dependent, but it is more common to adjust the wind speed, usually to plume height or 60% of the vertical spread, whichever is larger. Alberta data for long travel times show lateral plume growth at the 0.8 power of distance, whereas much world data reveals a near-linear growth. Wind shear seems to be responsible for maintaining a growth rate larger than the $\frac{1}{2}$ power predicted by the Taylor theory. The theory for concentration fluctuations is now well developed and can be used for predicting the probabilities of exceeding ensemble mean values.

Chemical transportation rates for sulphur dioxide in Alberta plumes are generally less than 0.5%/h at night and can get as large as 8%/h during summer days. Typical values appear to be 2-3% per hour. The conversion of nitric oxide to nitrogen dioxide is much more rapid, proceeding at rates of 0.2%/minute to 22%/minute. Dry deposition velocities for sulphur dioxide range from 0.25 cm/s over snow to 1.6 cm/s over ripe barley. Winter morning values are typically 0.6 cm/s and summer afternoon values are typically 1.2 cm/s.

4.17 REFERENCES

- Abel, K.H., J.A. Young, and L.A. Rancitelli. 1984. Atmospheric chemistry of inorganic emissions. Chap. 10. Atmospheric Science and Power Production. D. Randerson, ed. U.S. Dept. of Energy DOE/TFC-2760 (DE84005177). 423-465.
- Alberta Environment. 1988. User's Manual for Alberta Environment Dispersion models. Section 3 SULDEP3 (83 pp.) and Section 4 ADEPT (95 pp.). Air Quality Branch, Alberta Environment, Edmonton.
- Alp, E., M. J. E. Davies, R.G. Huget, L. H. Lam, and M.J. Zelensky. 1990. Gascon 2 A model to estimate ground-level H_2S and SO_2 concentrations and consequences from uncontrolled sour gas releases. prep. for Energy Resources Conservation Board by Concord Environmental Corporation. October 1990. ERCB Report 90-B, 275 pp.
- Angle, R.P. 1975. Airflow from mustard to fallow. *Atmosphere* 13, 110 - 125.
- Angle, R.P., H.S. Sandhu, and W.J. Schnitzler. 1979. Observed and predicted values of NO_2/NO_x in the exhaust plume from a compressor installation. *J. Air Poll. Contr. Assoc.* 29:253-255.
- Angle, R.P., C.S. Liu and K. Godard. 1987. Regulating oxides of nitrogen from compressor stations. Paper presented at CPANS/APCA Technical Meeting: New Directions in Air Quality Standards, Feb. 10, 1987, Calgary. 23 pp.
- Barr, S. and F.A. Gifford. 1987. The random force theory applied to regional scale tropospheric diffusion. *Atmos. Environ.* 21:1737 - 1741.
- Barrie, L.A. and J.L. Walmsley. 1978. A study of sulphur dioxide deposition velocities to snow in northern Canada. *Atmos. Environ.* 12:2321-2332.
- Barry, P.J. 1977. Stochastic properties of atmospheric diffusivity. Chapter 7 in Sulphur and its Inorganic Derivatives in the Canadian Environment. National Research Council, Ottawa, 313-358.
- Barry, P.J. 1971. A note on peak-to-mean concentration ratios. *Boundary Layer Meteor.* 2:122 - 126.
- Barry, P.J. 1969. Discussion. A statistical approach to the air pollution problem of chimney emission. *Atmospheric Environment* 3:328 - 330.
- Batchelor, G.K. 1949. Diffusion in a field of homogeneous turbulence. 1. Eulerian Analysis. *Australian. J. Sci. Res.* 2:437 - 450.
- Benjamin, S.F. 1975. The fate of SO_2 from sour gas plants. UNISUL-Group 4. September 1975. 33 pp.
- Benkley, C.W. and A. Bass. 1979. Development of mesoscale air quality simulation models. Volume 2, User's guide to MESOPLUME (mesoscale plume segment) model. NOAA contract no. 03-6-02-35254, Environmental Sciences Research Lab, Research Triangle Park, N.C. September 1979. 131 pp.

- Berger, A., J.L. Melice, and C.L. Demuth. 1982. Statistical distributions of daily and high atmospheric SO₂ concentrations. *Atmos. Environ.* 16, 2863 - 2877.
- Berkowicz, R., H.R. Olesen, and U. Torp. 1986. The Danish Gaussian air pollution model (OML); Description, test and sensitivity analysis in view of regulatory applications. *Air Pollution Modelling and its Application V.* C. De Wispalaere, F.A. Schiermeier and N.V. Gillani, Eds.. Plenum. New York, 453-481.
- Bigg, E. K., G.P. Ayers and D.E. Turvey. 1978. Measurement of the dispersion of a smoke plume at large distances from the source. *Atmos. Environ.* 12:1815 - 1818.
- Bottenheim, J.W. 1982. Plume chemistry studies during 1980 in the Athabasca sands area. Proc. symposium/workshop Acid Forming Emissions in Alberta and their Ecological Effects. Alberta Dept. of Environment, Canadian Petroleum Association, Oil Sands Environmental Study Group. 1982 March 9 to 12, Edmonton, 181-205.
- Bottenheim, J.W., and O.P. Strausz. 1982. Modelling study of a chemically reactive power plant plume. *Atmos. Environ.* 16:85 - 97.
- Bottenheim, J.W., and O.P. Strausz. 1981. Development of a chemically reactive plume model for application in the AOSERP study area. Prep. for the Alberta Oil Sands Environmental Research Program by Department of Chemistry, University of Alberta. AOSERP Report 117. 159 pp.
- Bottenheim, J.W., and O.P. Strausz. 1978. Computer modelling on polluted atmospheres and the conversion of atmospheric sulphur dioxide to sulphuric acid. Report to Alberta Environment, Project 76-43, June 1978. 144 pp.
- Bottenheim, J.W., and O.P. Strausz. 1977. Review of pollutant transformation processes relevant to the Alberta oil sands area. Prep. for the Alberta Oil Sands Environmental Research Program by the University of Alberta, Hydrocarbon Research Centre. AOSERP Report 25. 166 pp.
- Bowne, N.E. 1974. Diffusion rates. *J. Air Poll. Control Assoc.* 24:832 - 835.
- Briggs, G A. 1985. Analytical parameterizations of diffusion: the convective boundary layer. *J. Clim. Appl. Meteor.* 24:1167 - 1186.
- Briggs, G.A. 1973. Diffusion estimates for small emissions. Prelim. draft. Air Resources Turbulence and Diffusion Laboratory. Oak Ridge, Tennessee, ATDL contribution. File #79.
- Carpenter, S.B., T.L. Montgomery, J.M. Leavitt, W.C. Colbaugh and F.W. Thomas. 1971. Principal plume dispersion models. TVA power plants. *J. Air Poll. Control Assoc.* 21:491 - 495.
- Chadder, D.S., W.A. Murray, R.D. Brymer and A.H. Jamel. 1984. Improved measurement system for SO₂ dry deposition rates. Prepared for Alberta Environment by Promet Environmental Group Ltd. March 1984. 48 pp.
- Chamberlain, A.C. 1953. Aspects of travel and deposition of aerosol and vapour clouds. UKAEA Report No. AERE - HP/R-1261.

- Chatwin, P.C. and P.J. Sullivan. 1979. The core-bulk structure associated with diffusion clouds. Proc. Second Int. Symp. on Turbulent Shear Flows, London, July 1979.
- Cheng, L., A. Davis, E. Peake and D. Rogers. 1987. The use of aircraft measurements to determine transport, dispersion and transformation rates of pollutants emitted from oil sands extraction plants in Alberta. prep. for Alberta Environment, Research Management Division by Atmospheric Sciences Department, Alberta Research Council and Kananaskis Centre for Environmental Research, University of Calgary. AOSERP Report 133. 117 pp.
- Choukalos, M. 1980. A computer model of the risks from gas pipeline ruptures. Alberta Environment. December 1980. 43 pp.
- Choukalos, M. 1978. An integrated Gaussian puff atmospheric dispersion model. Alberta Environment. May 1978. 69 pp.
- Concord Scientific Corporation, 1986. Compressor station stack design Part I. Report to the Alberta Petroleum Industry Government Environmental Committee. Sept. 1986. 215 pp.
- Corbett, J.O. 1981. The validity of source-depletion and alternative approximation methods for a Gaussian plume subject to dry deposition. Atmos. Environ. 15:1207-1213.
- Cramer, H.E. 1976. Improved techniques for modelling the dispersion of tall stack plumes. Proc. 7th Int. Technical Meeting on Air Pollution Modelling and its Application. N. 51, NATO/CCMS. (NTIS PB 270 799). 731 - 780
- Cramer, H.E. 1957. A practical method for estimating the dispersal of atmospheric contaminants. Proc. First National Conf. on Applied Meteorology. American Meteorological Society, Hartford, Conn. Oct. 1957. C-33 to C-55.
- Csanady, G.T. 1973. Turbulent diffusion in the environment. D. Reidel Publishing Company, Boston. 248 pp.
- Csanady, G.T. 1969. Dosage probabilities and area coverage from instantaneous point sources on ground level. Atmos. Environ. 3:25 - 46.
- Daggupaty, S.M. 1980. A short range Gaussian puff dispersion model for accidental releases of radionuclides. First Mtg. of National Working Group on Atmospheric Radioisotope Releases, Atmos. Environ. Service, Downsview, Feb. 12, 1980. 53 pp.
- Davis, A., L. Cheng, and E. Peake. 1985a. Transformation of sulphur dioxide to particulate sulphate from oil sands extraction plant emissions. Paper presented at 19th Annual Congress of the Canadian Meteorological and Oceanographic Society, Montreal, Quebec. June 12-14, 1985. 25 pp.
- Davis, A., L. Cheng, and D. Rogers. 1985b. Winter observations of emissions from oil sands extraction plants. Paper presented at 19th Annual Congress of the Canadian Meteorological and Oceanographic Society, Montreal, Quebec. June 12-14, 1985.

- Davison, D.S. and E.D. Leavitt. 1979. Analysis of AOSERP plume sigma data. Prep. for the Alberta Oil Sands Environmental Research Program by Intera Environmental Consultants Ltd. AOSERP Report 63. 251 pp.
- Davison, D.S., C.J. Fortems, and K.L. Grandia. 1976. Plume dispersion measurements from an oil sands extraction plant, March 1976. prep. for the Alberta Oil Sands Environmental Research Program by Intera Environmental Consultants Ltd. AOSERP Report 13. 195 pp.
- DeMarrais, G.A. 1959. Wind-speed profiles at Brookhaven National Laboratory. *J. Meteor.* 16:181 - 190.
- Denison, P.J., T.A. McMahon and J.R. Kramer, 1979. Literature review on pollution deposition processes. Prepared for the Alberta Oil Sands Environmental Research Program and Syncrude Canada Ltd. by Acres Consulting Services Limited and Earth Science Consultants Inc. AOSERP Report 50. 264 pp.
- Djurfors, S. and D. Netterville. 1978. Buoyant plume rise in nonuniform wind conditions. *J. Air Poll. Control Assoc.* 28, 780 - 784.
- Doran, J.C., T.W. Horst and P.W. Nickola. 1978. Experimental observations of the dependence of lateral and vertical dispersion characteristics on source height. *Atmos. Environ.* 12:2259 - 2263.
- Draxler, R.R. 1984. Diffusion and transport experiments. Chap. 9. *Atmospheric Science and Power Production*. DOE/TIC-27601 (DE 84005 177), 367-422.
- Draxler R.R. 1979. Some observations of the along-wind dispersion parameter. Fourth Symp. on Turb. Diff. Air Poll., Jan. 15 - 18, 1979. Reno, Nevada. American Meteorological Society, Boston, 5-8.
- Draxler, R.R. 1976. Determination of atmospheric diffusion parameters. *Atmos. Environ.* 10:99 - 105.
- Egan, B.A. and A. Bass. 1976. Air quality modeling of effluent plumes in rough terrain. Proc. 3rd Symp. Atmos. Turb. Diff. and Air Quality. Raleigh, N.C. American Meteorological Society, 484-487.
- Eimutis, E.C. and M.G. Konicek. 1972. Derivations of continuous functions for the lateral and vertical atmospheric dispersion coefficients. *Atmospheric Environment*. 6:859 - 863.
- Energy Resources Conservation Board (ERCB). 1990. Field measurement program. Atmospheric tracer study under stable conditions and meteorological study. vol. 1. Oct. 1990. ERCB Report 90-B. Calgary, Alberta. 406 pp.
- Energy Resources Conservation Board (ERCB). 1979. User guide for multiple puff atmospheric dispersion model program No. PS10210. Energy Resources Conservation Board, Calgary, Alberta.

- Fanaki, F., R. Hoff, L. Barrie, R. Mickle, M. Lusi, K. Anlauf, A. Gallant, J. Kovalick, F. Froude, J. Markes, J. Arnold, S. Melnichuk, D. Brymer, A. Gaudenzi, A. Moser and D. Bagg. 1979a. Air system summer field study in the AOSERP study area, June 1977. prep. for the Alberta Oil Sands Environmental Research Program by Environment Canada, Atmospheric Environment Service. AOSERP Report 68. 248 pp.
- Fanaki, F., R. Mickle, M. Lusi, J. Kovalick, J. Markes, F. Froude, J. Arnold, A. Gallant, S. Melnichuk, D. Brymer, A. Gaudenzi, A. Moser and D. Bagg. 1979b. Air system winter field study in the AOSERP study area. February 1977. prep. for the Alberta Oil Sands Environmental Research Program by Fisheries and Environment Canada, Atmospheric Environment Service. AOSERP Report 24. 182 pp.
- Fanaki, F., R. Hoff and M. Lusi. 1979c. Air pollution studies at the Alberta oil sands area. PNWIS-APCA. Edmonton 1979. Paper 79-9.4. 15 pp.
- Fanaki, F., compiler. 1978. Meteorology and air quality winter field study in the AOSERP study area, March 1976. Prep. for the Alberta Oil Sands Environmental Research Program by Atmospheric Environment Service. AOSERP Report 27. 249 pp.
- Frenkiel, F.N., 1953. Turbulent diffusion - mean concentration distribution in a flow field of homogeneous turbulence. *Advan. Appl. Mech.* 3:61 - 107.
- Fung, C.S., E. Alp., and R.V. Portelli. 1985. Review and evaluation of Lagrangian models of long range transport and acidic deposition. Prep. for Canadian Electrical Association, Suite 580, One Westmount Square, Montreal, Quebec H3Z 2P9 by Concord Scientific Corporation. November 1985. 126 pp.
- Gifford, F.A. 1983. Atmospheric diffusion in the mesoscale range: the evidence of recent plume width measurement. Sixth Symp. on Turb. Diff. 22 - 25 March, 1983. American Meteor. Soc., Boston, MA, 300 - 304.
- Gifford, F.A. 1982. Long range plume dispersion. Comparison of the Mt. Isa data with theoretical and empirical formulas. *Atmos. Environ.* 16:883 - 886.
- Gifford, F.A. 1977. Tropospheric relative diffusion observations. *J. Appl. Meteor.* 16:311 - 313.
- Gifford, F.A. 1976. Turbulent diffusion typing schemes: a review, *Nuclear Safety.* 17:68 - 86.
- Gifford, F.A. 1975. Atmospheric dispersion models for environmental pollution applications. Chapter 2, Lectures on Air Pollution and Environmental Impact Analyses, American Meteorological Society, Boston, 35-58.
- Gifford, F.A. 1974a. The form of the frequency distribution of air pollution concentrations. Proc. Symp. on Statistical aspects of air quality data. Nov. 9-10, 1972 Chapel Hill, NC. US EPA - 650/4-74-038, 3-1 to 3-7.
- Gifford, F.A. 1974b. Discussion. Sensitivity of the Gaussian plume model. *Atmos. Environ.* 8:870 - 871.

- Gifford, F.A. 1968. An outline of theories of diffusion in the lower layers of the atmosphere. Chapter 3 of Meteorology and Atomic Energy. D. H. Slade, editor. U.S. AEC, available from NTIS as TID-24190, 65-116.
- Gifford, F.A. 1961. Use of routine meteorological observations for estimating atmospheric dispersion. *Nuclear Safety* 2(4):47 - 51.
- Gifford, F.A. 1960a. Atmospheric dispersion calculations using the generalized Gaussian plume model. *Nuclear Safety* 2(2):56 - 59.
- Gifford, F.A. 1960b. Peak to average concentration ratios according to a fluctuating plume dispersion model. *Int. J. Air Wat. Poll.* 3, 253 - 260.
- Gifford, F.A. 1959. Statistical properties of a fluctuating plume dispersion model. *Advances in Geophysics* 6:117 - 137.
- Gryning, S.E., A. A. M. Holtslag, J. S. Irwin and B. Sivertsen. 1987. Applied dispersion modelling based on meteorological scaling parameters. *Atmos. Environ.* 21:79 - 89.
- Hage, K.D., G. Arnason, N.E. Bowne, P.S. Brown, H.D. Entrekin, M. Levitz and J.A. Sekorski. 1966. Particle fallout and dispersion in the atmosphere. Final report. prep. by Travelers Research Center, Inc. for Sandia Corporation. SC-CR-66-2031.
- Hage, K.D. 1961a. On the dispersion of large particles from a 15-m source in the atmosphere. *J. Meteor.* 18:534-539.
- Hage, K.D. 1961b. The influence of size distribution on the ground deposit of large particles emitted from an elevated source. *Int. J. Air. Wat. Poll.* 4:24 - 32.
- Hales, J.M., D.C. Powell and T.D. Fox. 1977. STRAM - An air pollution model incorporating non-linear chemistry, variable trajectories and plume segment diffusion. Baitelle Pacific Northwest Lab. U.S. EPA Report 450/3-77/012. 157 pp.
- Hanna, S.R. 1986. Spectra of concentration fluctuations: the two time scales of a meandering plume. *Atmos. Environ.* 20:1131 - 1137.
- Hanna, S.R. 1984a. The exponential probability density function and concentration fluctuations in smoke plumes. *Boundary Layer Meteor.* 29:361 - 375.
- Hanna, S.R. 1984b. Concentration fluctuations in a smoke plume. *Atmos. Environ.* 18:1091 - 1106.
- Hanna, S.R. 1981. Effects of release height on σ_y and σ_z in daytime conditions. *Air Pollution Modelling and its Application I.* edited by C. De Wispelaere. Plenum Publishing, 337 - 356.
- Hanna, S.R., G.A. Briggs, J. Deardorff, B.A. Egan, F.A. Gifford and F. Pasquill. 1977. AMS workshop on stability classification schemes and sigma curves - summary of recommendations. *Bull. Amer. Meteor. Soc.* 58, 1305 - 1309.

- Hay, J.S. and F. Pasquill. 1959. Diffusion from a continuous source in relation to the spectrum and scale of turbulence. *Advances in Geophysics*. 6, Frenkiel and D. A. Shepphard (eds) Academic Press, N.Y., 345 - 365
- Hay, J.S. and F. Pasquill. 1957. Diffusion from a fixed source at a height of a few hundred feet in the atmosphere. *J. Fluid Mech.* 2 (Part 3):299 - 310.
- Heffter, J.L. 1965. The variation of horizontal diffusion parameters with time for travel periods of one hour or longer. *J. Appl. Meteor.* 4:153 - 156.
- Herman, M.N. 1980. Estimating long-term ground-level concentrations of SO₂ from short-term peak data. *J. Air Poll. Control Assoc.* 30:676 - 678.
- Hino, M. 1968. Maximum ground-level concentration and sampling time. *Atmos. Environ.* 2:149 - 165.
- Hoff, R.M. and F.A. Froude. 1979. Lidar observations of plume dispersion in northern Alberta. *Atmos. Environ.* 13:35 - 43.
- Högstrom, U. 1968. A statistical approach to the air pollution problem of chimney emission. *Atmos. Environ.* 2:251 - 271.
- Hogstrom, U. 1964. An experimental study on atmospheric diffusion. *Tellus XVI*: 205 - 251.
- Hollenbach, G.H. 1960. The estimation of ground concentrations of airborne material. Compiled by Imperial Oil Limited. A reissue of G.11737/Dt/10/58/25 as issued by the Air Ministry. 12 pp.
- Hosker, R.P. Jr. 1984. Flow and diffusion near obstacles. Chap. 7. *Atmospheric Science and Power Production*. D. Randerson, editor. USDOE DOE/TIC-27601 (DE 8400177). 241-326.
- Hosker, R.P., Jr. 1973. Estimates of dry deposition and plume depletion over forests and grassland. Presented at the IAEA Symp. on the Physical Behavior of Radioactive Contaminants in the Atmosphere. Vienna, Nov. 12-16, 1973. IAEA-SM-181/19. ATDL Contribution File No. 85. Air Resources Atmospheric Turbulence and Diffusion Lab., Oak Ridge, Tennessee. 27 pp.
- Hosker, R.P. and S.E. Lindberg, 1982. Review: Atmospheric deposition and plant assimilation of gases and particles. *Atmos. Environ.* 16, 889-910.
- Horst, T.W. 1978. Comments on "A general gaussian diffusion-deposition model for elevated point sources." *J. Appl. Meteor.* 17:415 - 416.
- Horst, T.W. 1977. A surface depletion model for deposition from a Gaussian plume. *Atmos. Environ.* 11:41 - 46.
- Hunt, J.E., T. Low, W. A. Murray and D. Chadder. 1981. An analysis of techniques for measuring the dry deposition rate of SO₂. Prep. for Alberta Environment by Concord Scientific Corporation and Promet Environmental Group. 83 pp.

- Ide, Y. 1971. The effect of the observation time on turbulent diffusion in the atmosphere. *Wat. Air Soil Poll.* 1:32 - 41.
- International Atomic Energy Agency (IAEA). 1976. Safety guide on meteorology - climatology, diffusion and transport in nuclear power plant siting. 67 pp.
- Irwin, J.S. 1983. Estimating plume dispersion - a comparison of several sigma schemes. *J. Climate and Appl. Meteor.* 22:92 - 114.
- Islitzer, N.F. and D.H. Slade. 1968. Diffusion and transport experiments, Chapter 4 of Meteorology and Atomic Energy. D. H. Slade (editor) U.S. AEC, 117 - 188.
- Johnson, O. 1967. Comparison of observed and predicted dosages of phosgene from instantaneous sources. Suffield memorandum No. 48/67. DRB Project D52-32-01-10. Defence Research Establishment Suffield, Ralston Alberta. 11 pp.
- Kumar, A. 1979a. The effects of cross-wind shear on horizontal dispersion. paper 79 - 29.3. 72nd Annual Mtg. Air Poll. Contr. Assoc. Cincinnati, Ohio. June 24 - 29, 1979. 16 pp.
- Kumar, A. 1979b. Estimation of atmospheric dispersion coefficients for elevated releases. Fourth Symp. Turb. Diff. Air Poll. Jan. 15 - 18, 1979. Reno, NV. American Meteorological Society, Boston, 19 - 26.
- Kumar, A. and S. Djurfors. 1977. General expressions for maximum and critical ground level concentrations. Syncrude Canada Ltd., Professional Paper 1977-6. 22 pp. (Presented as: Kumar, A. and S. Djurfors. 1978. Maximum and critical ground level concentrations from a synthetic crude oil plant. Paper No. 78-48.5. 71st Annual Meeting of the Air Pollution Control Association, Houston, Texas, 25 - 30 June 1978).
- Lamb, R.O. 1979. The effects of release height on material dispersion in the convective planetary boundary layer. Fourth Symposium on Turbulence, Diffusion and Air Pollution. Reno, NV, Jan. 15 - 18, 1979. Amer. Meteor. Soc., Boston. 25-33.
- Lamb, R.G. 1968. An air pollution model of Los Angeles. M. S. Thesis, University of California, Los Angeles.
- Leahey, D.M. and B. Burns 1979. Estimates of sulphur deposition in the vicinity of Alberta's sulphur extraction plants. Paper 79-3.3. PNWIS-APCA Meeting, Edmonton 1979.
- Leahey, D.M., M.J.E. Davies and H.A. Bambrough. 1981. Values of NO_2/NO_x observed in the exhaust plume from a turbine driven compressor installation. Paper presented at the 1981 Annual Meeting of the Pacific Northwest International Section of the Air Pollution Control Association. 16 pp.
- Leahey, D.M., M.J.E. Davies and H.A. Bambrough. 1983. Values of NO_2/NO_x observed in the exhaust plume from a turbine driven compressor installation. *Atmos. Environ.* 17(5): 991-997.

- Leahey, D.M. and M.C. Hansen. 1985. Meteorological evidence for greater dispersion of elevated plumes. *J. Air. Poll. Contr. Assoc.* 35: 828 - 830.
- Lee, G. K., H. Whaley and J.G. Gainer. 1973. Plume dispersion at sulphur plants. Energy Processing Canada, November/December; 24-34.
- Legge, A.H. and S.V. Krupa (eds) 1990. *Acidic Deposition: Sulphur and Nitrogen Oxides*. Lewis Publishers. 659 pp.
- Legge, A.H. 1988. The present and potential effects of acidic and acidifying air pollutant on Alberta and Environment. Critical Point 1. Final Report. Prep. for Acid Deposition Research Program by Kananaskis Centre for Environmental Research, University of Calgary. ADRP-B-16-88. 79 pp.
- Leung, T.B. 1972. Sulphur plant stack dispersion. M. Sc. Thesis, University of Calgary, Dept. of Chemical Engineering. July 1972. 83 pp.
- Leung, T., M.F. Mohtadi and R.D. Rowe. 1972. Evaluation of stack gas sizing formula. Gas Processing Canada. May - June 1972, 40-42.
- Lewellen, W.S. and M. Teske. 1975. Turbulence modeling and its application to atmospheric diffusion. EPA Report. EPA600/4-75-016a,b. part 1, 90 pp. part 2, 59 pp.
- Lin, C.C. and W. H. Reid. 1963. Turbulent flow, theoretical aspects. *Handbuch der Physik*. Vol. 8, Part 2. Springer Verlag, Berlin. 438 - 523
- Lloyd, A.C. and F.W. Lurmann. 1985. Chemical transformations in air quality models. Chap. 28. *Handbook of Applied Meteorology*. David D. Houghton, ed. John Wiley and Sons, New York. 767-777.
- Lott, R.A. 1977. A quasi-study technique for modeling the atmospheric dispersion of continuous plumes. Tennessee Valley Authority, Air Quality Branch. Muscle Shoals, AL. 50 pp.
- Ludwig, F.L. 1984. Transport and diffusion calculations during time-and space-varying meteorological conditions using a microcomputer. *J. Air Poll. Contrd. Assoc.* 34:227 - 232.
- Ludwig, F.L., L.S. Gasiorok, and R.E. Ruff. 1977. Simplification of a Gaussian puff model for real-time minicomputer use. *Atmos. Environ.* 11:431 - 436.
- Lusis, M.A., K.G. Anlauf, L.A. Barrie and H.A. Wiebe. 1978. Plume Chemistry studies at a northern Alberta power plant. *Atmos. Environ.* 12:2429 - 2437.
- Martin, D.O. and J.A. Tikvart. 1968. A general atmospheric diffusion model for estimating the effects on air quality of one or more sources. Presented at 61st Annual Meeting 9 of the Air Poll. Control Assoc. June 23-27, Saint Paul, MN. paper 68-148.
- McElroy, J.L. and F. Pooler. 1968. St. Louis Dispersion Study. Vol. II Analysis. US DHEW Report AP-53.

- McMahon, T.A. and P.J. Denison, 1979. Empirical atmospheric deposition parameters - A survey. *Atmos. Environ.* 13:571 - 585.
- Meade, D.J. 1960. Meteorological aspects of the safety and location of reactor plants. WMO Tech. Not. No. 33. 13 - 22.
- Melli, P. and E. Runca. 1979. Gaussian plume model parameters for ground-level and elevated sources derived from the atmospheric diffusion equation in a neutral case. *J. Appl. Meteor.* 18:1216 - 1221.
- Millan, M. and Y. S. Chung. 1977. Detection of a plume 400 km from the source. *Atmos. Environ.* 11:939 - 944.
- Miller, C.W. 1978. An examination of Gaussian plume dispersion parameters for rough terrain. *Atmos. Environ.* 12:1359 - 1364.
- Montgomery, T.L. and J.H. Coleman. 1975. Empirical relationships between time-averaged SO₂ concentrations. *Environ-Sci. and Tech.* 9:953 - 957.
- Montgomery, T.L., S.B. Carpenter, and H.E. Lindley. 1971. The relationship between peak and mean SO₂ concentrations. Conf. on Air Poll. Meteor, April 5 - 9, 1971. Raleigh, N.C. Amer. Meteor. Soc., Boston, 136 - 137.
- Moore, D.J. 1974. Observed and calculated magnitudes and distances of maximum ground level concentration of gaseous effluent material downwind of a tall stack. *Advances in Geophysics* 18B, Academic Press. New York, 201 - 221.
- Mueller, S.F. and L.R. Reisinger. 1986. Measured plume width versus sampling time; a look beyond 10 kilometers. *Atmos. Environ.* 20:895 - 900.
- Murray, W.A. 1980. New approaches to stack design. *Energy Processing Canada*, January - February 1980:48 - 49.
- Nester, K., W.G. Hubschmann and P. Thomas. 1977. The influence of ground roughness on atmospheric diffusion. *Proc. 4th International Clean Air Congress*. May 16-20, Tokyo, 153 - 156.
- Netterville, D.D.J. 1979. Concentration fluctuations in plumes. *Environmental Research Monograph* 1979-4. Syncrude Canada Ltd., Edmonton, 228 pp.
- Netterville, D.D.J. and D.J. Wilson. 1980. Probability estimates of high concentrations in plumes. *Fifth International Clean Air Congress*, Buenos Aires, Argentina. October 20 - 25, 1980, 599-607.
- Nickola, P.W. 1976. Measured differences in real and apparent plume parameters. *Third Symp. Atm. Turb. Diff. and Air Quality*, Oct. 12 - 22, 1976 Raleigh, N.C. Amer. Meteor. Cos. Boston MA, 283 - 287.
- Nickola, P.W. 1971. Measurements of the movement, concentration and dimensions of clouds resulting from instantaneous point sources. *J. Appl. Meteor.* 10:562 - 973.

- Overcamp, T.J. 1978. Reply. *J. Appl. Meteor.* 17:417 - 418.
- Overcamp, T.J. 1976. A general Gaussian diffusion-deposition model for elevated point sources. *J. Appl. Meteor.* 15:1167-1171.
- Pasquill, F. 1976. Atmospheric dispersion parameters in Gaussian plume modelling, Possible requirements for change in the Turner workbook values. part II. EPA-600/4-76-030b, 53 pp.
- Pasquill, F. 1975. The dispersion of material in the atmospheric boundary layer - the basis for generalization. Chapter I, Lectures on Air Poll. and Environmental Impact Analyses. American Meteorological Society. Boston. 1 - 34.
- Pasquill, F. 1974. Atmospheric Diffusion, 2nd edition. Halsted Press. John Wiley & Sons, New York. 429 pp.
- Pasquill, F. 1971. Atmospheric dispersion of pollution. *Quart. J. Roy. Meteor. Soc.* 97:369 - 395.
- Pasquill, F. 1968. Discussion. The determination of stack height. *Atmos. Environ.* 2:228.
- Pasquill, F. 1962. Atmospheric Diffusion. 1st edition. D. Van Nostrand Company Ltd., London. 297 pp.
- Pasquill, F. 1961. The estimation of the dispersion of windborne material. *Met. Mag.*, 90:33 - 49.
- Pasquill, F. and F.B. Smith. 1983. Atmospheric Diffusion. 3rd ed. Ellis Horwood Ltd. Halsted Press: a division of John Wiley & Sons, Chichester, 437 pp.
- Pasquill, F. and F.B. Smith. 1971. The physical and meteorological basis for the estimation of the dispersion of windborne material. *Proc. 2nd Int. Clean Air Congress.* Dec. 6-11, 1970. Washington. Academic Press, 1067 - 1072
- Pendergrass, W. and S.P.S. Arya. 1984a. Dispersion in neutral boundary layer over a step change in surface roughness - I Mean flow and turbulence structure. *Atmos. Environ.* 18:1267 - 1279.
- Pendergrass, W. and S.P.S. Arya. 1984b. Dispersion in neutral boundary layer over a step change in surface roughness - II concentration profiles and dispersion parameters. *Atmos. Environ.* 18:1281 - 1296.
- Petersen, W.B. 1986. A demonstration of INPUFF with the MATS data base. *Atmos. Environ.* 20:1341 - 1346.
- Peterson, K.R. 1968. Continuous point source plume behavior out to 160 miles. *J. Appl. Meteor.* 7:217 - 226.
- Plate, E. ed. 1982. Engineering Meteorology. Elsevier Scientific Publishing Company, Amsterdam, 740 pp.

- Ragland, K.W. 1976. Worst case ambient air concentrations from point sources using the Gaussian plume model. *Atmos. Environ.* 10:371 - 374.
- Ramsdell, J.V. Jr. and W.T. Hinds. 1971. Concentration fluctuations and peak-to-mean concentration ratios in plumes from a ground-level continuous point source. *Atmos. Environ.* 5, 483 - 495.
- Randerson, D. 1984. Atmospheric boundary layer. Chap. 5. Atmospheric Science and Power Production. U. S. Dept. of Energy. DOE/TIC-27601 (DE84005177), 147 - 188.
- Randerson, D. 1972. Temporal changes in horizontal diffusion parameters of a single nuclear debris cloud. *J. Appl. Meteor.* 11:670 - 673.
- Reid, J.D. 1981. The appropriate choice of σ_z and u for predicting surface concentration using the Gaussian model. *Proc. Air Pollution Sessions. CMOS 15th Annual Congress, Saskatoon, Sask. May 27 - 29, 1981*, 207 - 216.
- Reid, J.D. 1979. Markov chain simulation of vertical dispersion in the neutral surface layer for surface and elevated releases. *Boundary Layer Meteorology.* 16:3 - 22.
- Roberts, O.F.T. 1923. The theoretical scattering of smoke in a turbulent atmosphere. *Proc. Roy. Soc. London A*, 104:640 - 654.
- Roberts, J. J., E. J. Croke, A. S. Kennedy, J. E. Norco and L. A. Conley. 1970a. A multiple-source urban atmospheric dispersion model. Argonne National Laboratory Center for Environmental Studies. Publication ANL/ES-CC-007. 148 pp.
- Roberts, J.J., E.J. Croke, and A.S. Kennedy. 1970b. An urban atmospheric dispersion model in *Proc. symp. on multiple source urban diffusion models*. A. C. Stern, editor. US EPA, Air Pollution Control Office Publication AP-86, Research Triangle Park, North Carolina, 6-1 to 6-72.
- Roberts. J.J., J.E. Norco, A.S. Kennedy, and E.J. Croke. 1970c. A model for simulation of air pollution transients. *Second International Clean Air Congress. Dec. 6-11, 1970. Washington*, 1161-1167.
- Rowe, R.D. 1981. Comments on "lateral dispersion coefficients as functions of averaging time". *J. Appl. Meteor.* 20:728 - 730.
- Rowe, R.D. 1975. Discussion. The derivation of plume dispersion parameters from measured three-dimensional data. *Atmos. Environ.* 9:462.
- Rowe, R.D., M.F. Mohtadi, J.J. Havlena, D.I. Exall, J.M. Boyle and S.F. Benjamin, 1977. The short range chemistry of sour gas plant plumes. *Proceedings of Alberta Sulphur Gas Research Workshop III, Edmonton, Alberta, November 17-18, 1977. Alberta Department of the Environment, Research Secretariat, Edmonton*, 27 - 38.
- Sagendorf, J.F. and C.R. Dickson. 1976. Diffusion under low windspeed, inversion conditions. *Third Symposium on Atmospheric Turbulence, Diffusion and Air Quality. Oct. 19 - 22, 1976. Raleigh, N. C. Amer. Meteor. Soc., Boston*, 277-282.

- Sakiyama, S.K. and P.A. Davis. 1988. Additional field verification of convective scaling for the lateral dispersion parameter. *J. Appl. Meteor.* 27:882 - 884.
- Sakiyama, S.K., P.F. Freeman, and R.P. Angle. 1988. Dispersion model for use in the design of compressor station stacks - Paper presented at the APCA - CPANS Annual Meeting, May 16 - 17, 1988, Calgary, Alberta. 20 pp.
- Sakuraba, S., M. Moriguchi and I. Yamazi. 1973. The splitting of plume centreline height in elevated-source diffusion. *Papers in Meteorology and Geophysics.* 24:13 - 30.
- Sakuraba, S., M. Moriguchi and I. Yamazi. 1972. The elevated source diffusion at coastal areas. *Papers in Meteorology and Geophysics,* 27:307 - 328.
- Sakuraba, S., Moriguchi and I. Yamazi. 1969. A comparative study of surface source dispersion and elevated-source dispersion. *Papers in Meteorology and Geophysics.* 20:337 - 364.
- Sandhu, H.S., H.P. Sims, R.A. Hursey, W.R. MacDonald and B.R. Hammond, 1980. Environmental sulphur research in Alberta: A review. Research Secretariat. Alberta Environment. Feb. 1980. 96 pp.
- Scire, J.S., F.W. Lurmann, A. Bass, and S.R. Hanna. 1984. User's guide to the MESOPUFF II model and related processor programs. EPA-600/8-84-013. USEPA. prep. by Environmental Research and Technology Inc., Concord, MA. Research Triangle Park. N.C. (NTIS PB 89-181755). 226 pp.
- Scorer, R.S. 1968. *Air Pollution.* Pergamon Press, Oxford. 151 pp.
- Sehmel, G.A. 1980. Particle and gas dry deposition: A review *Atmos. Environ.* 14:983 - 1011.
- Seinfeld, J.H. 1986. *Atmospheric Chemistry and Physics of Air Pollution.* John Wiley and Sons, New York. 738 pp.
- Senshu, T., S. Yshida and K. Nakamura. 1974. A model of the boundary layer for the prediction of atmospheric pollution. *Symp. Atmospheric Diffusion and Air Pollution.* Sept. 9 - 13, 1974, Santa Barbara, American Meteorological Society, Boston, 180 - 183.
- Shieh, C.M. 1978. A puff pollutant dispersion model with wind shear and dynamic plume rise. *Atmos. Environ.* 12:1933 - 1938.
- Shieh, L. 1969. A multiple source model of turbulent diffusion and dispersion in urban atmospheres. Report No. NR-69-11. Geophysical Sciences Laboratory, New York University, December 1969.
- Shieh, L.J., B. Davidson and J.P. Friend. 1970. A model of diffusion in urban atmospheres: SO₂ in greater New York. *Proc. Symp. on multiple source urban diffusion models.* editor. A. C. Stern. Air Pollution Control Office Publ. AP-86, E. P. A. Research Triangle Park, North Carolina, 10-1 to 10-39.

- Singer, I.A. and M.E. Smith. 1966. Atmospheric dispersion at Brook haven National Laboratory. *Int. J. Air and Wat. Poll.* 10:125 - 135.
- Singer, I.A., K. Imai, and R.G. Del Campo. 1963. Peak to mean pollutant concentration ratios for various terrain and vegetation cover. *J. Air Poll. Control Assoc.* 13:40 - 47.
- Slade, D. 1965. Dispersion estimates from pollutant releases of a few seconds to 8 hours in duration. Tech. Note 2-ARL-1, Air Resources Laboratory, Washington, D.C. 23 pp.
- Slawson, P.R., G. Davidson, W. McCormick and G. Raithby. 1978. A study of dispersion characteristics of the G.C.O.S. plume. Prep. by Envirodyne Limited for Syncrude Canada Ltd. April 1978. 113 pp.
- Smith, F.B. 1972. A scheme for estimating the vertical dispersion of plumes from a source near ground level. *Proc. Third Meeting Expert Panel on Air Pollution Modelling. NATO Committee on Challenges of Modern Society.* XVII - 1 to XVII - 14.
- Smith, M.E. (editor). 1968. Recommended guide for the prediction of the dispersion of airborne effluents. The American Society of Mechanical Engineers. New York. 85 pp.
- Smith, M.E. and I.A. Singer. 1966. An improved method of estimating concentrations and related phenomena from a point source emission. *J. Appl. Meteor.* 5:631 - 639.
- Stewart, R.E. and G.T. Csanady. 1967. Deposition of heavy particles from a continuous elevated source. *Proc. 1st Canadian Conference on Micrometeorology, Part II*, 395.
- Strom, G.H. 1976. Transport and diffusion of stack effluents. *Air Pollution*, Vol. 1, 3rd ed. A.C. Stern (ed) Academic Press, 401 - 501.
- Strom, G.H. 1968. Atmospheric dispersion of stack effluents. Chapter 8, *Air Pollution*, vol. 1, 2nd ed. A.C. Stern (ed) Academic Press, 227 - 274.
- Sutton, O.G. 1953. *Micrometeorology*. McGraw-Hill, New York. 333 pp.
- Sutton, O.G. 1932. A theory of eddy diffusion in the atmosphere. *Proc. Roy. Soc. (London), Ser. A.* Vol. 135:143 - 165.
- Tadmor, J. and Y. Gur. 1969. Analytical expressions for the vertical and lateral dispersion coefficients in atmospheric diffusion. *Atmospheric Environment* 3:688 - 689.
- Taylor, G.I. 1921. Diffusion by continuous movements. *Proc. London Math Soc.* Series 2.20, 196 - 212.
- Taylor, G.I. 1915. Eddy motion in the atmosphere. *Phil. Trans. R. Soc.* 215A:1 - 26.
- Turner, D.B., 1970. Workbook of atmospheric dispersion estimates. Office of Air Programs Publ. No. AP-26, USEPA, Research Triangle Park, North Carolina. 84 pp.

- Ukeguchi, N., H. Okamoto and Y. Ide. 1973. Predictions of calm inversion pollution. Proceedings of the 3rd International Clean Air Congress. Oct. 8-12. Dusseldorf, B66 - B68.
- Van der Hoven, I. 1976. A survey of field measurements of atmospheric diffusion under low-wind-speed inversion conditions. Nuclear Safety 17(2):223 - 230.
- Van der Hoven, I. 1972. Southwest Energy Study. prep. by Southwest Energy Federal Task Force. Appendix E. Meteorology. 186 pp.
- Venkatram, A., D. Strimaitis and D. Dicristofaro. 1984. A semiempirical model to estimate vertical dispersion of elevated releases in the stable boundary layer. Atmos. Environ. 18, 923 - 928.
- Vet, R.J., N.W. Reid, E. Alp, and S.J. Diehl. 1982. Compressor exhaust chemistry and wake diffusion. Prep. for Alberta Petroleum Industry Government Environmental Committee by Concord Scientific Corporation. Report No. 67-1. April 1982. 205 pp.
- Vogt, K.J. 1977. A new system of release height dependent diffusion parameters for the Gaussian plume model. Proc. The Fourth International Clean Air Congress, May 16-20, Tokyo, 171 - 175.
- Voldner, E.C., L.A. Barrie, and A. Sirois, 1986. A literature review of dry deposition of oxides of sulphur and nitrogen with emphasis on long-range transport modelling in North America. Atmos. Environ. 20:2101 - 2123.
- Walker, E.R. 1965. A particulate diffusion experiment. J. of Appl. Meteor. 4:614 - 621.
- Weber, A.H. 1976. Atmospheric dispersion parameters in Gaussian plume modelling Part I. review of current systems and possible future developments. U. S. EPA, EPA/600/4-86/030a. 74 pp.
- Weil, J.C. 1985. Updating applied diffusion models. J. Climate and Appl. Meteor. 24:1111 - 1130.
- Weil, J.C. and R.P. Brower. 1984. An updated Gaussian plume model for tall stacks. J. Air Poll. Control Assoc. 34:818 - 827.
- Weil, J.C. and R.P. Brower. 1982. The Maryland PPSP dispersion model for tall stacks. Ref. No. PPSP-MP-36; Maryland Power Plant Siting Program. Martin Marietta Corp. Batimore.
- Western Research and Development. 1978. Derivation of first order estimates of sulphur deposition in the region of representative point sources. Prep. for Alberta Environment. November 13, 1978. Ref.: 2737-R1. 61 pp.
- Whaley, H. and G. K. Lee. 1982. The behavior of buoyant plumes in neutral and stable conditions in Canada. Atmos. Environ. 16(11):2555 - 2564.
- Whaley, H., G.K. Lee and J.G. Gainer. 1980. The behavior of buoyant merging plumes in the Rocky Mountain Foothills. J. Air Poll. Cont. Assoc. 30:782 - 786.

- Whaley, H. and G.K. Lee. 1978. An assessment of plume dispersion parameters measured in fall and winter at a tar-sands refinery complex. *J. Air Poll. Control Assoc.* 28, 589 - 593.
- Whaley, H. 1974. The derivation of plume dispersion parameters from measured three-dimensional data. *Atmos. Environ.* 8:281 - 290.
- Wilson, D.J. 1987. Defining effective turbulence velocities for dispersion in a stable atmosphere. prep. for Energy Resources Conservation Board. ERCB Report 90-B, vol. 7, Appendix D. 9 pp.
- Wilson, D.J. 1986. Plume dynamics and concentration fluctuations in gas emissions. prep. for Alberta Environment by Dept. of Mechanical Engineering, University of Alberta, Edmonton. 99 pp.
- Wilson, D.J. 1982. Predicting risk of exposure to peak concentrations in fluctuating plumes. Prepared for Alberta Environment by the Mechanical Engineering Dept., University of Alberta, Edmonton. 90 pp.
- Wilson, D.J. 1981. Along-wind diffusion of source transients. *Atmos. Environ.* 15:489 - 495.
- Wilson, D.J. 1979a. Wind tunnel simulation of plume dispersion at Syncrude Mildred Lake site. prep. by Mech. Eng. Dept. Univ. of Alberta for Syncrude Canada Ltd. Syncrude Environmental Research Monograph 1979 - 1, Edmonton. 198 pp.
- Wilson, D.J. 1979b. The release and dispersion of gas from pipeline ruptures. Prep. by Dept. of Mech. Eng., University of Alberta for Alberta Environment. January 1979. 91 pp.
- Wilson, D.J. and B. Bara. 1985. Crosswind profiles of conditionally averaged mean concentration in plumes. 7th Symp. Turb. & Diff., Boulder, CO, Amer. Meteor. Soc. Nov. 1985, 259 - 261.
- Wilson, D.J., A.G. Robins and J.E. Fackrell. 1985. Intermittency and conditionally-averaged concentration fluctuation statistics in plumes. *Atmos. Environ.* 19:1053 - 1064.
- Wilson, D.J. and B.W. Simms. 1985. Exposure time effects on concentration fluctuations in plumes. Prep. for Alberta Environment by Dept. of Mechanical Engineering, University of Alberta, Edmonton. 160 pp.
- Wilson, D.J., A.G. Robins and J.E. Fackrell. 1982a. Predicting the spatial distribution of concentration fluctuations from a ground level source. *Atmos. Environ.* 16:497 - 504.
- Wilson, D.J., A.G. Robins and J.E. Fackrell. 1982b. Concentration fluctuations in an elevated plume: a diffusion - dissipation approximation. *Atmos. Environ.* 16:2581 - 2589.
- Wilson, R.B., G.E. Start, C.R. Dickson and N.R. Ricks. 1976. Diffusion under low wind speed conditions near Oak Ridge, Tennessee. Third Symposium on Atmospheric Turbulence Diffusion and Air Quality. Oct. 19 - 22, 1970. Raleigh, N.C. Amer. Meteor. Soc., Boston, 269 - 276.

- Yamartino, R.J. 1985. Atmospheric pollutant deposition modelling. Chap. 27. Handbook of Applied Meteorology. D.D. Houghton (ed), John Wiley and Sons, 754 - 766.
- Yamartino, R.J. 1977. A dispersion model for low wind speed conditions: theory and experimental application. Paper 77-58.6. 70th Annual Meeting of the Air Poll. Cont. Assoc. Toronto, June 20 - 24, 1977.
- Zannetti, P. 1986. A new mixed segment-puff approach for dispersion modelling. Atmos. Environ. 20:1121 - 1130.
- Zannetti, P. 1981. An improved puff algorithm for plume dispersion simulation. J. Appl. Meteor. 20:1203 - 1211.

5.0 FLOW AND DIFFUSION OVER RUGGED TERRAIN

Alberta is topographically diverse with extreme variations of terrain -- wide expanses of prairies, rolling foothills and towering mountains. Industrial development is not limited to topographically simple regions and thus the issue of pollutant transport and diffusion in complex or rugged terrain is of practical concern.

Topographic features affect both the plume trajectory and the rate of diffusion. The airflows can be very complex since they are a result of the combined effects of locally induced circulations and incident flow disturbances. The shape of terrain features together with their thermal and roughness characteristics control plume behavior in concert with atmospheric stability. This added level of complexity brings added complexity to the modelling of such processes.

5.1 WIND AND TURBULENCE EFFECTS

The flow features unique to regions of complex terrain can be classified into four causal categories. They can be due to interactions of terrain with synoptically driven flows (large scale flows driven by horizontal pressure gradients), locally driven flows, interactions between these two flow regimes, and other secondary effects (Dickerson 1980).

5.1.1 Synoptically Driven Flows

5.1.1.1 Channelling. For near-neutral stability, large-scale winds moving at an oblique angle to a valley or set of ridges can result in winds parallel to the valley or ridges. This is commonly referred to as channelling. Channelling can also refer to the increase in wind speeds which occur when wind is forced through a narrow opening (such as a mountain pass). Examples of this behavior are shown in Figures 5.1a and 5.1b.

Channelling (where the large-scale winds have been deflected and result in winds parallel to the ridges) can occur even with rather gentle terrain features. Leahey and Hansen (1982) analyzed 2300 pilot balloon observations near the Suncor oil sands plant to reveal channelling of the synoptic-scale winds in the broad Athabasca River valley. The valley has a broad plain 50 km wide and is bounded by the Birch and Muskeg mountains (450 and 300 m above the research facility elevation respectively).

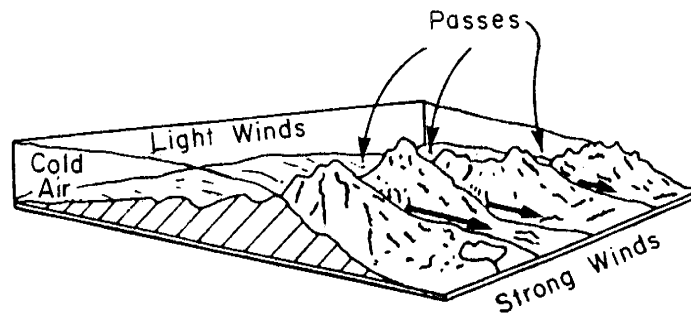


Figure 5.1a Wind speed-up through mountain pass (Orgill et al. 1971).

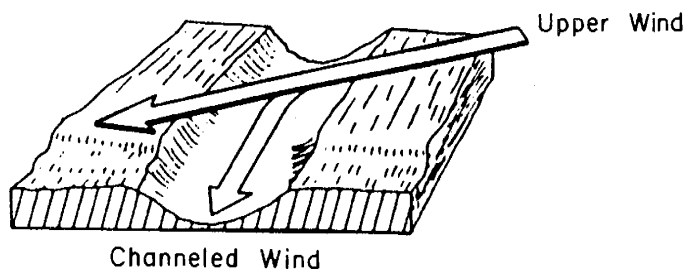


Figure 5.1b Wind direction change due to presence of valley (Orgill et al. 1971).

5.1.1.2 Wakes. As high Reynolds number flows approach an obstacle a region of modified turbulence occurs in the lee of the obstacle. This region is defined as a *wake* and is a result of random vortex shedding off the obstacle and subsequent cascading to smaller and smaller scales. Wakes can occur on large scales where they can dominate some features of the turbulence for long distances (see Scorer and Verkaik 1989 for satellite photographs of turbulence in the lee of mountain ranges and islands).

Yahalom (1972) studied the flow separation and wake phenomena in a University of Calgary wind tunnel modelling study using a scale model of the Bow River valley west of Calgary and incident westerly flow. A recirculation region (cavity) of high turbulence and low wind speed extended downwind 10 - 15 ridge heights and upward to 2-2.5 ridge heights. Beyond the cavity the wake grew wider and deeper but the velocity turbulence difference from the incident flow diminished.

An increase in the turbulence of the incident flow reduces the extent to which the wake is distinguishable (Halitsky et al. 1965). A 10% velocity defect in the wake may be detectable as far as 60 ridge heights downwind (Arya and Shipman 1979).

5.1.1.3 Flow as it Encounters an Isolated Hill. As flow approaches a hill, a *speed-up* occurs as the flow moves over the crest. This behavior was evident over Kettles Hill (near Pincher Creek, Alberta - elliptical shape, 100 m high and a slope of about 0.08 along the longitudinal axis). The speed-up phenomenon was studied extensively at this site using kites, tethersondes and a series of surface tower measurements and the results are reported in Taylor et al. (1983a) and Mickle et al. (1984). They found that under neutral conditions, wind speeds at the hilltop were more than 60% greater than those measured upwind, a surprising amount considering the low profile of this hill.

When the atmosphere is stably stratified, a number of phenomena can result when the synoptic-scale flow encounters an obstacle. As the flow encounters a ridge and rises up over it, a *lee wave* may form (Figure 5.2). This wave will have a certain frequency and amplitude and can extend large distances downwind. A standing lee wave may be evidenced by longitudinal cloud rolls or lenticular clouds at the wave crests.

Lester and MacPherson (1976) and Lester (1976a) reported observations of waves in the lee of the Rockies in southern Alberta during a chinook event. The chinook arch cloud was embedded in the crest of a gravity wave with amplitude of 800 m and vertical motions of 1.8 ms^{-1} . Lester (1976b) developed a climatology of wave clouds from 12 years of observations made at Pincher Creek, Alberta. Wave clouds occurred 141 days per year with a distinct October-March maximum and summer minimum in frequency of occurrence.

It is also possible that as the flow moves over a ridge, it accelerates on the lee side. As the accelerating flow is forced to slow down, a *hydraulic jump* can occur. The occurrence of a hydraulic jump can be determined on the basis of a *Froude number* (a measure of the relative magnitude of inertial and buoyant forces in a fluid). This jump (Figure 5.3) is characterized by local pressure changes, turbulence and occasionally, by convective cloudiness. Similar findings have been obtained in laboratory situations (Snyder and Hunt 1984).

If the fluid has insufficient kinetic energy to move up and over the obstacle, then the flow will be *blocked*. A stagnation zone develops upstream of the obstacle or, in the case of an isolated hill, the fluid moves horizontally around the object. This type of flow condition leads to adverse air quality since a plume can actually impinge on the hillside or enter a stagnation zone and accumulate pollutants. The situation is shown in Figures 5.4b, and c and is contrasted to the flow field experienced in neutral conditions (5.4a).

An observational study of this type of flow over a ridge in stable conditions was undertaken in a series of tetroon (constant-level balloon) releases, reported in Chung and Rowe (1979), and plume photographs at Copithorne Ridge, southwest of Calgary. Benjamin et al. (1977, 1978) used smoke trails to visualize the flow over the ridge and found that, for stable conditions, the ridge induced horizontal (directional) shear. This indicates that plume impingement could occur since the occurrence of horizontal shear is consistent with flow attempting to go around ridge rather than over it. Later photographic studies of a cooling tower plume as it passed over this ridge are reported by Lee (1979) and Rowe et al. (1980). The trajectories were consistent with a model which assumes that the flow above a certain height above ground moves over the ridge, and below this height moves around it.

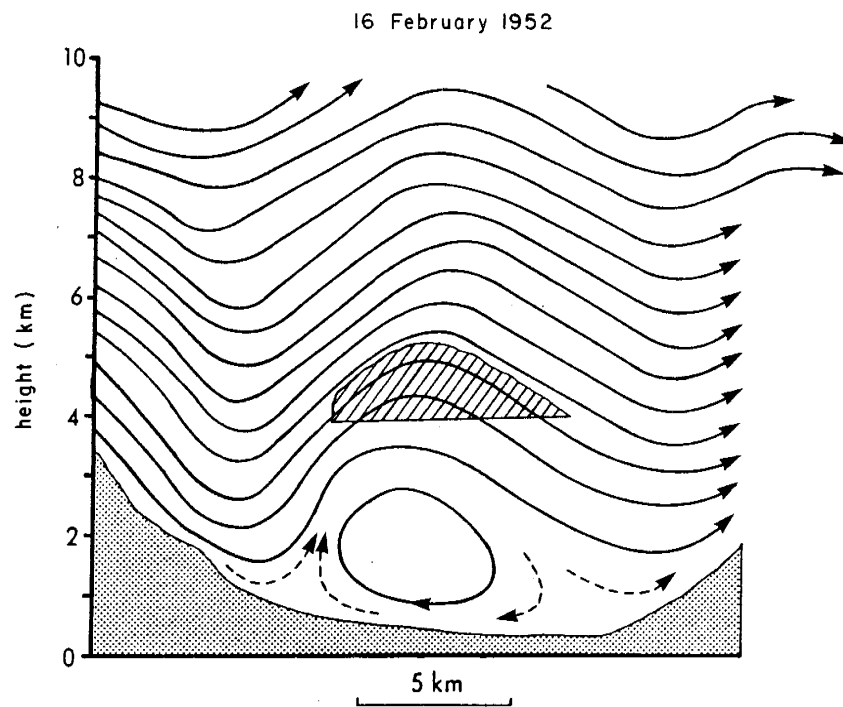


Figure 5.2 Lee waves, cloud and rotor flow to the lee of Sierra Nevada Mountains (Atkinson 1981).

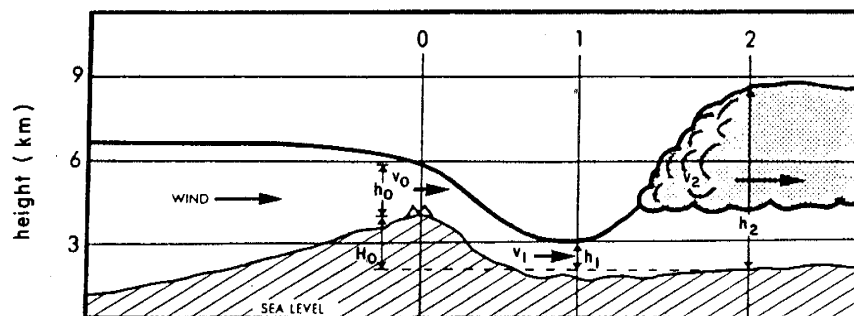


Figure 5.3 Hydraulic jump (Atkinson 1981).

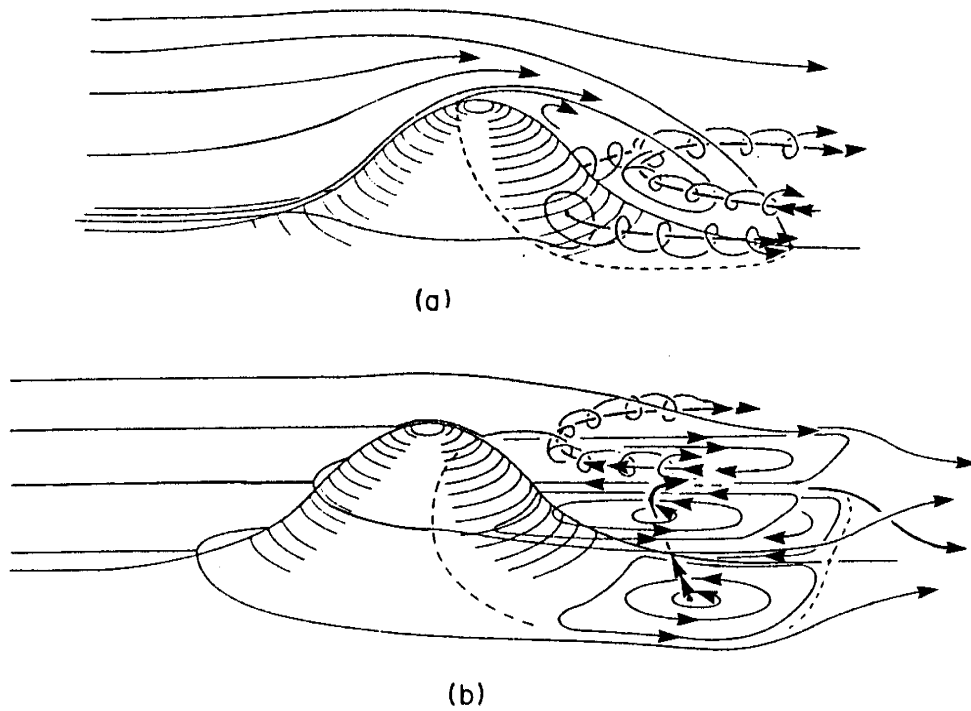


Figure 5.4 Flow past a three-dimensional bell-shaped hill for a) neutral and b) highly stratified flows. Dashed line shows regions of separated flow. (Hunt et al. 1979).

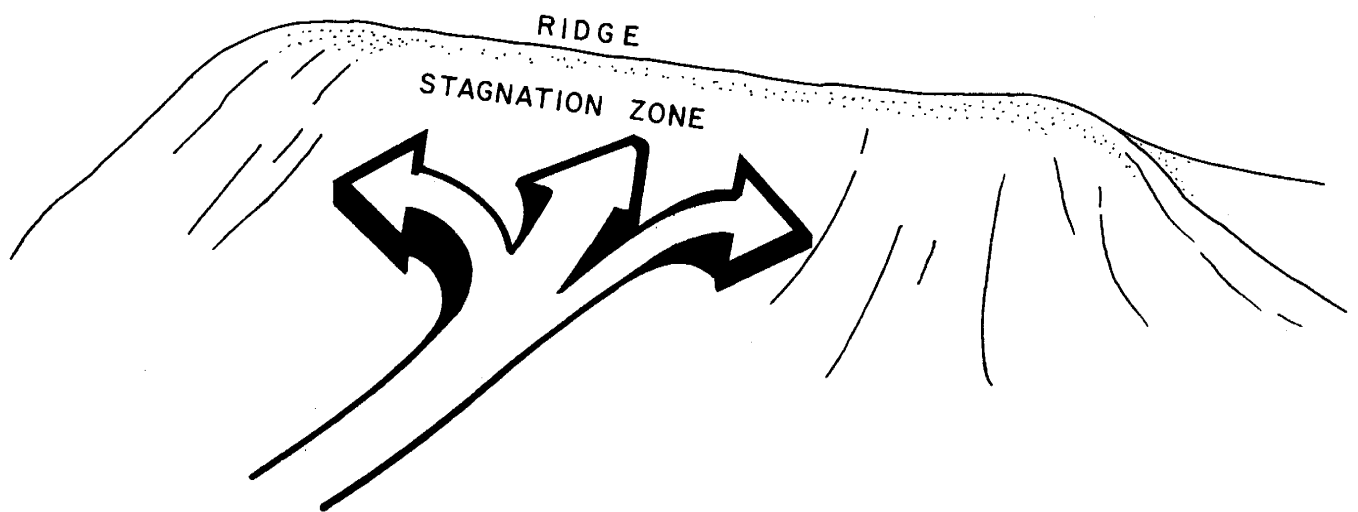


Figure 5.4c Blocking and stagnation flow.

Flow over irregular terrain in both neutral/unstable and stable conditions has been observed over a wide range of scales and terrain types (Nicholls 1973; Mason and Sykes 1979; Bradley 1980; Taylor et al. 1983a, b; Strimaitis 1983).

5.1.1.4 Implications for Plume Transport and Diffusion. The streamlines indicating flow over a terrain feature in Figures 5.4a converge as they approach and pass over the feature. From continuity principles, this convergence results in a speed-up of the fluid. A plume embedded in the flow field is brought closer to the surface and undergoes volume stretching. In addition, the plume depth is compressed due to streamline convergence and the plume width is extended (if you vertically squash a plume, it grows horizontally). This effect is most pronounced under stable conditions. Finally, turbulence near these features is enhanced. The plume reflects this turbulence enhancement by increased plume spreads. Under stable conditions the plume may impinge on the side of the hill and attempt to go around the obstacle rather than over it (Figure 5.4b).

Wakes also affect pollutant concentrations when the plume becomes embedded in one. The increased turbulence enhances plume spread, and the decreased wind speeds decrease plume stretching. Some of these effects serve to increase ground-level concentrations; others lead to a decrease. This will be discussed further in the section on dispersion modelling in complex terrain.

5.1.2 Locally Driven Flows

Some air motions in complex terrain are generated by forcing mechanisms of purely local origin and can behave independently of the synoptic-scale winds. The forcing mechanisms are thermal in nature, where surface heating or cooling results in local pressure gradients.

During a clear night with light synoptic-scale winds, long-wave radiative cooling results in a temperature inversion. If the surface is sloped, the near-surface air, which is cooler and more dense, will move downhill. The behavior is analogous to water running downhill. These winds are called *drainage* or *katabatic* winds. Drainage winds are a regular occurrence during clear nights with in most mountainous regions of the world (Defant 1951, Bergen 1969), and have been documented photographically (Thorpe and

Orgill 1986). Longley (1968, 1969) showed that the wind directions measured at Calgary International Airport had a diurnal behavior which indicated drainage flows off the eastern slopes of the Rocky Mountains.

The velocity of such katabatic winds can be small, often less than 1 m/s. Observations are thus made difficult and the air current is easily disturbed by surface obstructions. Low-lying areas are prone to an accumulation of cool air either because of a depression with no exit or because obstacles such as trees act as a "cold-air dam." The resulting "cold pools" have greatly intensified nocturnal inversions. (The term "frost hollow" describes an agricultural outcome of drainage flows). On a larger scale, the drainage of air from glaciers, plateaus or highlands can result in devastating winds that exceed 30 m/s.

Another type of locally generated flow is called an *upslope* or *anabatic* wind. During the day when the slope is heated by the sun, the air near the surface is heated and rises. The translation of upward motion into upslope flow is thought to occur because of the chimney effect at ridge top. Convection at the top of the ridge produces vertical motions and cumulus buildups. Longley (1968, 1969) showed that afternoon wind directions at Calgary are indicative of flow up the eastern slopes of the Rocky Mountains. Thyer (1981) confirmed this behavior by an examination of pilot balloon flights at Edmonton and Calgary, where vector average winds for two summer months were used to delineate the vertical variation of winds. Thyer hypothesized that the low-level wind behavior as evidenced in the profiles at Calgary is a result of heating on the eastern slopes of the Rocky Mountains which induces a large-scale upslope wind, modified in direction by the Coriolis force.

The drainage and upslope flow phenomena are illustrated in Figure 5.5.

5.1.2.1 Valley Systems. The upslope/downslope flow phenomena help to explain the flow patterns experienced in a valley. Under downslope flow conditions, air moves down the side walls and fills the valley with a down-valley flow. Under upslope flow conditions this situation is reversed. Thus under clear sky conditions, where radiative heating and cooling effects are the greatest, the valley flow system exhibits different characteristics during different times of the day (Defant 1949) as shown in Figures 5.6a and 5.6b.

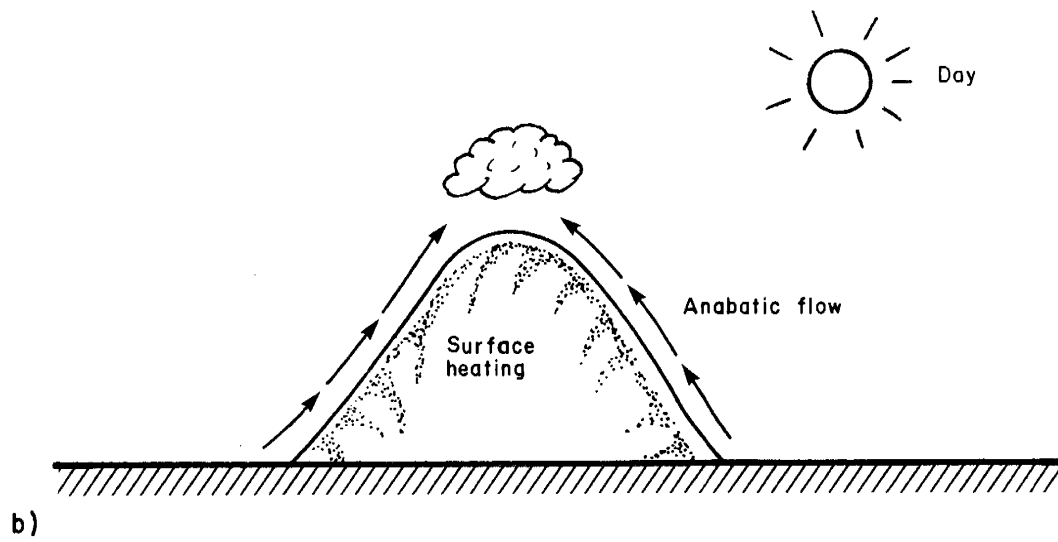
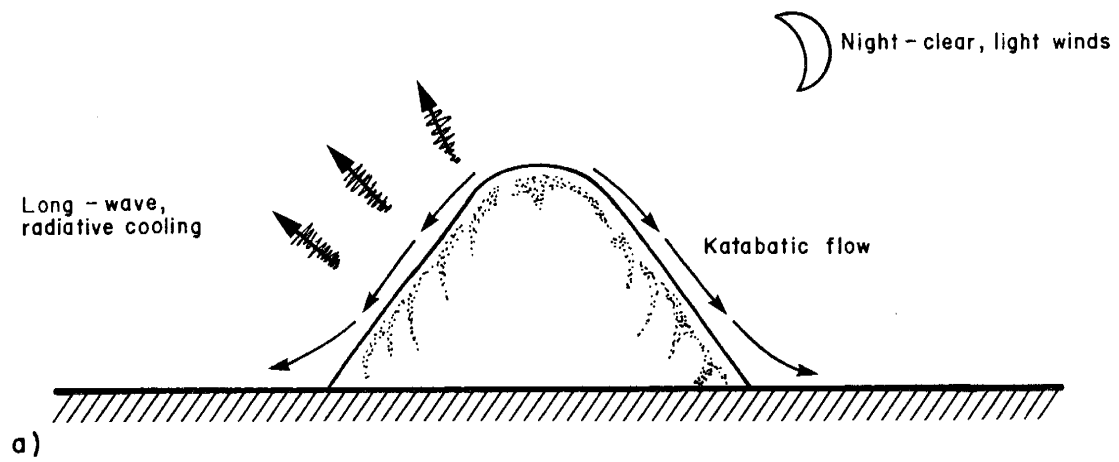


Figure 5.5 a) Katabatic (downslope) and b) anabatic (upslope) flows and their associated conditions.

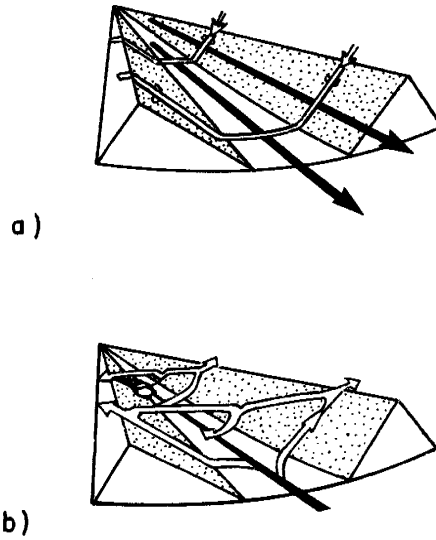


Figure 5.6 Valley circulations under clear skies and weak regional winds for a) night and b) early afternoon. (Defant 1949).

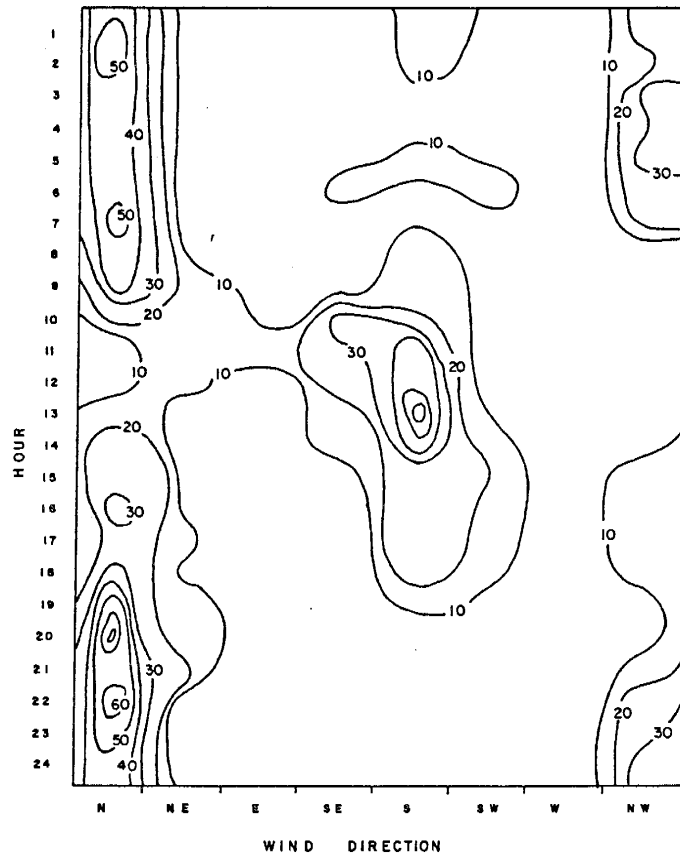


Figure 5.7 Hourly percent frequency distribution of wind direction in Canyon Creek valley, Sept. 1982 (Angle and Gourlay 1983).
Note: Flows down-valley are north, up-valley are south.

Up-valley and down-valley flow behavior has been observed to occur in a variety of different valley types around the world (e.g., Vermont - Davidson and Rao 1963; California - Gudiksen et al. 1983; Colorado - Whiteman 1982; Austria - Freytag 1987). Table 5.1 lists the studies which confirm the up-valley/down-valley flow behavior in a variety of valley types and locations in Alberta. An example of the diurnal influence on up- and down-valley phenomena is shown in Figure 5.7 which summarizes surface wind data collected over 1 month during the summer in the Elbow River valley, (a broad open valley 210 m high and 3800 m wide) located west of Bragg Creek, Alberta. The plots show a high frequency of down-valley winds during the night and the reverse situation during the day. The up-valley wind frequency is lower than the down-valley wind frequency because of synoptic interference; that is, afternoon convective activity causes a coupling between the surface and synoptic winds. This behavior is also evident in the MacHattie (1968) analysis (see Table 5.1 for description) for a deep Kananaskis valley and in other locations in the world (Manins and Sawford 1979 - broad flat valley in Australia).

Down-valley flow is characterized by a drainage "jet" where the down-valley wind speed reaches a maximum part way up the valley depth. These distinctive nose-shaped profiles have been observed in a variety of valleys (Clements et al. 1989). The maximum wind speed can occur at half the valley depth (Davidson and Rao 1963, Munn 1966); other observational studies have shown that other ratios exist. Wind speed and direction profiles from the Canyon Creek valley in Alberta during nocturnal drainage flow conditions as reported by Angle and Gourlay (1983), Gourlay et al. (1984) and Sakiyama (1990) are shown in Figure 5.8a and b.

The drainage flow ceases after the sun begins to warm the slopes and valley floor. This results in convective mixing which breaks up the temperature inversion and induces upslope valley flows. The breakup of inversion/drainage flow typically occurs 3-5 h after theoretical sunrise (Whiteman 1982).

Table 5.2 summarizes the results of Alberta studies which report these characteristics.

Table 5.1 Observations of up- and down-valley flow behavior in Alberta valleys.

Reference	Valley Characterization	Comments
Klassen (1962)	North Saskatchewan River valley at Edmonton, 1-1.5 km wide, average depth about 50 m	drainage flows with (maximum speed 2.5 m/s) drainage interrupted when winds above valley > 4 m/s, no up-valley winds observed valley slope wind 2.5 m deep
Paterson and Hage (1979)	same valley as above	down-valley wind occurs when prevailing wind is cross-valley valley slope wind attains maximum depth of 15 m
MacHattie (1968)	Kananaskis valley (valley bounded by Mt. Lorette, Mt. Allan and Mt. Kidd on the west and Mt. McDougall on the east)	hourly surface observations during July/Aug. show distinct down-valley flow during the night and up-valley during the day anabatic flow suffered some synoptic-scale interference
Leahey and Hansen (1982)	Athabasca River Valley, north of Fort McMurray, 50 km wide bounded by Birch and Muskeg Mtns (450 and 300 m high respectively)	pilot balloon showed katabatic influences on winds 400 m above ground
Angle and Gourlay (1983) Gourlay et al. (1984)	Canyon Creek valley (180 m deep, 1120 m wide) and Elbow River valley (210 m deep, 3800 m wide) both valleys are west of Bragg Creek, Alberta	surface wind observations show classic up- and down-valley behavior in both valleys

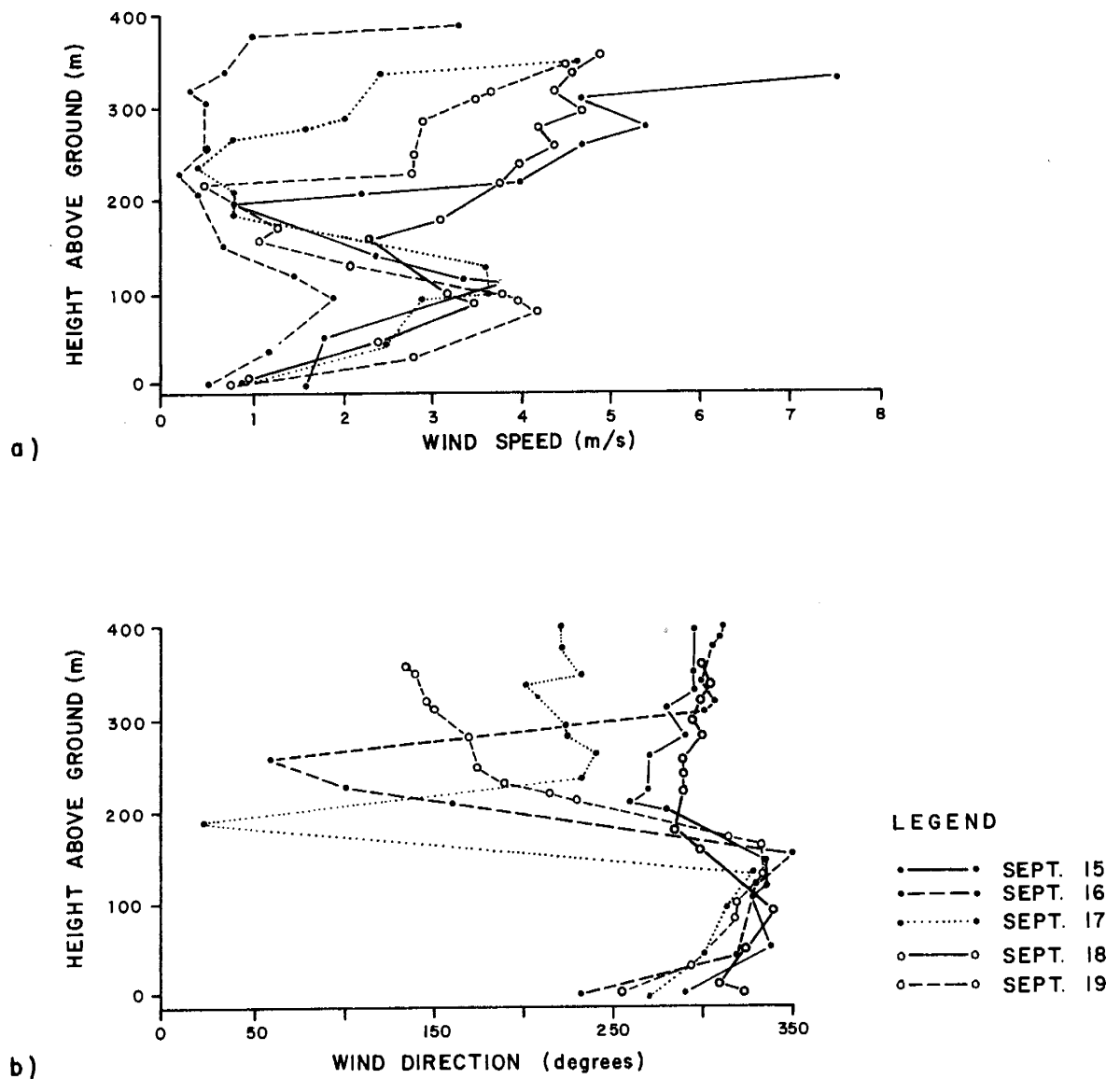


Figure 5.8

a) Wind speed and b) direction profiles in Canyon Creek valley at 06:00 MST during clear weather (Angle and Gourlay 1983). Note: Down-valley direction is 345° .

Table 5.2 Nocturnal valley flow characteristics from observational evidence in Alberta.

Reference	Valley Description	Height of Maximum Wind Speed (times the valley depth)	Inversion/ Drainage Total Breakup (hours after sunrise)	Down-valley Wind Speed Maximum (m/s)
Klassen (1962)	North Saskatchewan River valley at Edmonton 1-1.5 km wide, average depth 50 m	0.3 to 0.4	2 to 6	2.5
Angle and Gourlay (1983) Gourlay et al. (1984)	Elbow River valley, west of Bragg Creek 210 m deep, 3800 m wide	0.5	3 to 5	6.9
Sakiyama (1990) Sakiyama et al. (1984)	Canyon Creek valley west of Bragg Creek 180 m deep (east ridge), 1130 m wide	0.45 for synoptic winds with up-valley component, 0.55 when this is not the case	3 to 4	4.9

5.1.2.2 Turbulence in Nocturnal Drainage Flow. Elevated turbulence levels are a common feature in valley drainage flow. The sources of this turbulence are many and include:

- (1) drainage flow interaction with valley walls, outcroppings and rough valley floors (Start et al. 1973, 1975),
- (2) shear-generated turbulence from the nocturnal jet,
- (3) interactions of the main drainage jet with tributary flow (Porch et al. 1989),
- (4) disturbances of the drainage flow by the regional winds (Barr and Orgill 1989), and
- (5) the existence of cellular circulations which result in surges (Stone and Hoard 1989).

Angle and Gourlay (1983) reported turbulence measurements in the Elbow River valley during clear nights where inversions were present and well developed drainage flow occurred. They found that turbulence levels were larger than what would be found under similar conditions in flat terrain. This result was reflected in the diffusion study of Sakiyama (1983) and Sakiyama and Angle (1988) which measured ground-level concentrations of a tracer gas, sulphur hexafluoride, released into the Elbow River valley, nocturnal drainage flow. Dilution was much greater in stable drainage flows (equivalent to neutral flows) than it would be under stable conditions in flat terrain.

5.1.2.3 Morning Transition. Between a period of fully developed nocturnal drainage flow and a period of up-valley flow, a complex flow adjustment occurs as the valley atmosphere changes in accordance with the change in energy loss/input over a diurnal cycle. Studies in different valleys around the world show a common morning transition behavior where inversion breakup is characterized by the upward growth of the daytime convective boundary layer and the descent of the top of the inversion. This latter feature does not occur in flat terrain.

This type of inversion breakup behavior has been observed in Switzerland (Muller and Whiteman 1988) and Colorado (Whiteman 1982). Figure 5.9 shows the inversion descent and breakup behavior observed in the Canyon Creek valley near Bragg Creek, Alberta reported in Sakiyama et al. (1984) and Sakiyama (1990). One mechanism proposed to explain the descent of the inversion top during inversion breakup is based on mass continuity (Whiteman 1982). The top descends to compensate for the mass evacuated from the central portion of the valley by the thermally generated upslope flows blowing up the valley sidewalls. In other words, the principle of mass continuity requires that the inversion descend. This can have important implications for valley plume behavior (Whiteman and McKee 1978) which will be discussed later. Another proposed mechanism is a scouring of the cold stable core from above by turbulent mixing at the top of the drainage layer by the above-valley winds (Lenshow et al. 1979 and Davidson and Rao 1963). It is also possible that, if the cold air which feeds the drainage is cut off (which occurs at sunrise), the cold air will eventually drain out the end of the valley and the top of the inversion will descend (Ayer 1961). This is analogous to the drop in the river level that occurs when the feeding streams are cut off.

5.1.2.4 Implications for Plume Transport and Diffusion. The significance of valley flows for the dispersal of chemical plumes is often misunderstood. Although valleys are favorable locations for the formation of inversions, turbulence actually increases because of the katabatic flow (Leahey and Halitsky 1973, Sakiyama 1983, Sakiyama and Angle 1988). The main problem is associated with the preferred wind directions as indicated by a marked elongation of the wind rose along the valley axis. This means that the same areas will be exposed to plume material over and over again and inversion breakup fumigations are likely to occur with regularity over much the same receptor area (Hewson and Gill 1944). Long-term average concentrations would thus be substantially higher than if the chemical source were located on the plain. Turbulence levels are highly variable and can lead to the sudden appearance of plume material in high concentrations. Finally, the valley walls limit lateral plume spread which prevents the usual reduction of concentration with distance. If vertical exchange above the valley is also limited, the valley becomes a tube filled with a uniform concentration of pollutants, much higher than would be experienced over an open plain.

If the inversion descent behavior is due to the subsidence of the stable core (Whiteman 1982), a plume in this stable core will be brought downward as the core

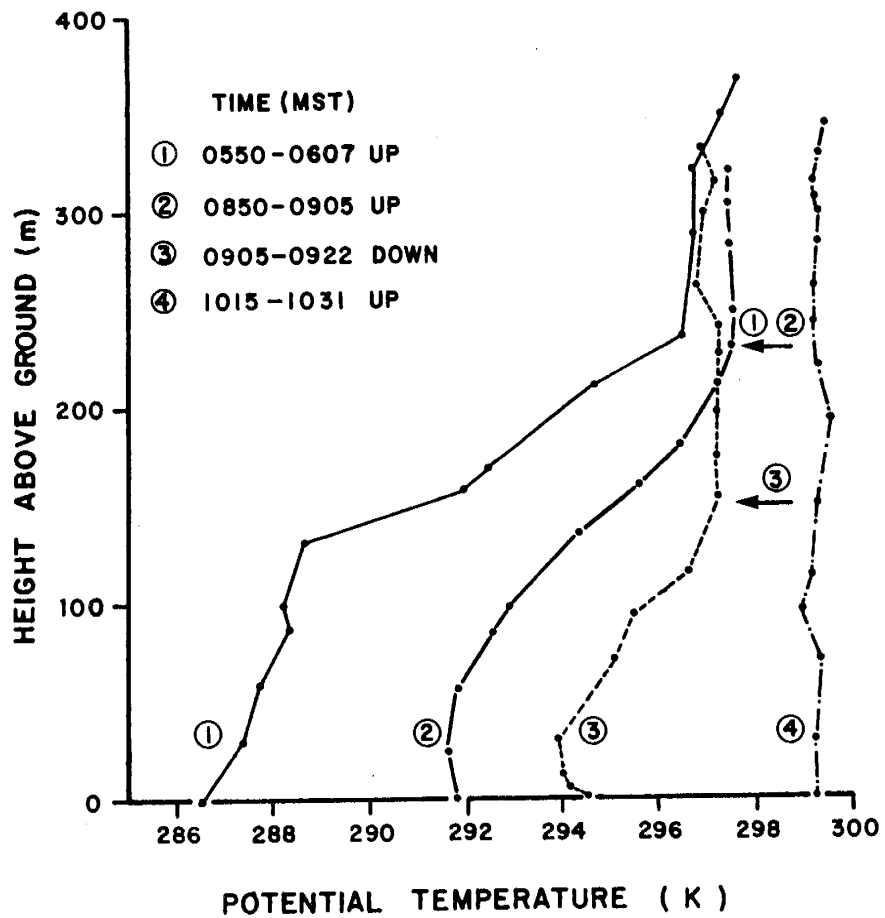


Figure 5.9 Time series of potential temperature profiles taken in the Canyon Creek valley on September 19, 1982. The top of the inversion is indicated by arrows (Sakiyama 1990).

descends. The descent will continue until pollutants are entrained in the convective boundary layer at the valley bottom or shallow slope flow layer (Whiteman and McKee 1978). After being entrained into this layer the pollutants are removed from the valley by advection of the upslope layer. This process is dramatically different from the behavior of a plume in flat terrain during inversion breakup and can result in high concentrations at the valley bottom and sidewalls.

5.1.3 Interactions Between Flow Systems

The characteristics of the wind and turbulence fields in complex terrain can be determined from interactions between different flow systems. The resulting new flow region may not be of purely local origin or synoptically driven. For example, when the turbulent wakes from adjacent peaks overlap, a new flow region is established in the overlap region even though the individual wakes result from synoptically driven flows.

There are several examples where synoptic flow may interact with local flow. The up-valley wind flows in the deep Kananaskis valley studied by McHattie (1968) and the Canyon Creek and Elbow River valleys studied by Angle and Gourlay (1983) showed an overriding of the local wind system by the synoptic flow. The convective activity in the valley caused a coupling between the local and synoptic flows with the resulting loss of up-valley flow. It is also entirely possible that the synoptic flow can enhance the up-valley flow if it is generated (or channelled) in the same direction. In addition, it is also possible that the synoptic flows can be so strong they can disrupt the nocturnal drainage flow (Neff and King 1987) or even remove the air as a coherent mass, especially in broad, open valleys (Lenschow et al. 1979).

An example of interactions between local flows is the situation where the drainage flow from a feeder valley joins the main down-valley drainage flow. Klassen (1962) found that the drainage from a steep narrow ravine entering the North Saskatchewan River valley at Edmonton, could be detected over 100 m past the ravine mouth in the river valley. In addition, observations of vertical motions over the river indicated the existence of a cellular structure superimposed on the drainage flow. This cellular structure may be driven by the heat from the river (Munn 1966). This is an example of two superimposed local flows resulting in a new flow regime.

5.1.3.1 Implications for Plume Transport and Diffusion. As a general rule, the zones of interaction between two flow regions, whether they are of synoptic or of local origin, results in enhanced turbulence or flow deformations. This, in turn, results in an increase in mixing and a decrease in concentration. For example, observations of the Suncor powerhouse plume reported by Slawson et al. (1978) revealed an interaction zone between the valley drainage flow (the Athabasca River valley is 50 m deep and 2 km wide) and the synoptic flow. The interaction zone caused dramatic cross-wind plume spreads due primarily to flow deformation.

However, the interactions may also cause a change in the path of a plume, for instance, due to the deflection caused by tributary drainage flow entering the main flow. This may result in high concentrations in areas where none were experienced before.

The effects of flow interaction on chemical transport and diffusion depend on the individual situation. It is difficult to determine what will happen to a plume in flows that are of purely synoptic or local origin; it is even more formidable to determine plume behavior when these two flow regions interact with each other or themselves.

5.1.4 Secondary Effects

5.1.4.1 Flow Separation and Recirculation Rotors. An airstream approaching an abrupt rise in terrain elevation (cliff, bluff, plateau or tableland) in neutral or unstable stratification will be deflected upward near the edge. If the rise is steep, then a *bolster eddy* may form at the base. This occurs because the streamlines cannot follow the sharp bend, or in other words, because an adverse pressure gradient brings the incident flow to rest and then reverses it. If the edge is rounded, the flow accelerates smoothly onto the higher terrain. If the edge is sharp, separation may occur, giving rise to a zone of recirculation and a wake of enhanced turbulence. An intermediate possibility is that the streamlines curve over the edge and back to a bluff, producing a sheltered region of very light winds. This behavior is illustrated in Figure 5.10.

Prominent ridges, in neutral conditions, cause flow separation near the apex. Yahalom (1972) observed the separation phenomenon in a University of Calgary wind tunnel modelling study using a scale model of the Bow River valley west of Calgary and

an incident westerly flow. The lee side of the ridge was occupied by a recirculation zone of high turbulence and low mean speed known by various names as the *cavity*, *bubble*, *eddy* or *stagnation region*. The cavity extended downwind about 10 - 15 ridge heights and upward to 2 - 2.5 ridge heights.

When the wind blows perpendicular to a valley under neutral and unstable conditions, separation at the ridge top may fill the valley with a *recirculation eddy* (Figure 5.11). Angle and Gourlay (1983) reported the occurrence of easterly and northeasterly winds in the Canyon Creek valley (east of Bragg Creek, Alberta) with a westerly regional flow, suggestive of a recirculation eddy. This behavior may also explain the high frequency of east winds observed in this valley reported by Webb (1982). A helical recirculation eddy has been postulated for Colorado's Brush Creek valley (Barr and Orgill 1989). Wind profiles during nighttime, down-valley flow show winds at the valley bottom oriented in the opposite direction to the regional cross-valley flow. A stagnant zone at about half the valley depth corresponds to the centre of the eddy.

5.1.4.2 Flow Oscillations. Slope winds and valley winds are generally intermittent (Munn 1966). Steady-state conditions are rarely maintained for any length of time. Drainage winds usually occur in surges rather than continuously and complete flow reversals are not uncommon. A range of pulse lengths have been observed in various valley drainage flows around the world (Pack Canyon, Colorado, 15 minutes - Coulter et al. 1988; Greenland, 45 minutes - Gyning et al. 1985; 90 minutes - Doran and Horst 1981). The circulation is in a delicate balance, easily disrupted by gravity waves, roughness changes, flow confluences or blocking by obstacles. The down-valley direction may not be the same as that of water flow; if the valley is constricted, the cold air may flow upstream towards where the valley broadens (Munn 1966).

Hage (1981) found wave-like oscillations in wind speed during drainage flow conditions in the North Saskatchewan River valley at Edmonton. The observed period ranged from 16 - 33 minutes and the pulses were probably due to standing internal gravity waves present in a small river basin upstream from the observation site.

5.1.4.3 Compensation Airflows. Compensation flows are generally required by mass continuity and circulation cells. In valleys, compensation airflows are called "anti-wind systems" because they have been observed immediately above the valley drainage flow

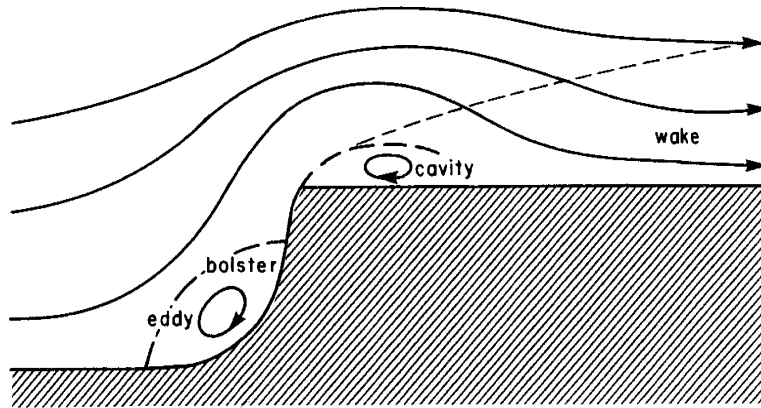


Figure 5.10 Airflow toward a steep bluff.

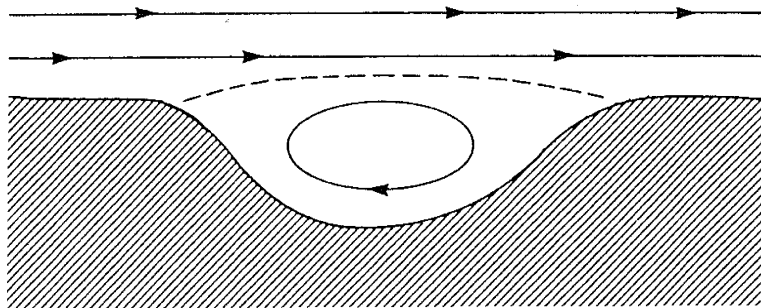


Figure 5.11 Separation in strong cross-valley winds may fill the valley with one large eddy.

(Buettner and Thyer 1966) and are oriented in the up-valley direction. They are postulated to exist from mass continuity considerations – air flowing down-valley is replaced by compensatory flow above the drainage flow.

The "anti-wind system" is not a common feature of valley wind systems. Sakiyama (1990) found no evidence for this type of compensation flow in the Canyon Creek valley. An alternate hypothesis (Whiteman and Barr 1986) postulated that the air flowing down the valley is replaced by air from above. That is, a general subsidence is occurring over the valley. Rates of 0.1 ms^{-1} for the Brush Creek valley in Colorado were calculated from mass conservation principles.

Evidence for cellular compensation flow is reported in Klassen (1962) and Hage (1980, 1979) for the North Saskatchewan River valley at Edmonton. Vertical motions over the center of the valley are compensated by flow from the valley sides, resulting in two counter-rotating cells superimposed on the valley drainage flow.

5.1.4.4 Implications for Plume Transport and Diffusion. Secondary flow effects can have a minor or dramatic effect on chemical dispersal depending on the scale of the phenomena relative to the plume size. Separation cavities can be so large that the entire plume can get trapped, resulting in high concentrations in the cavity region (see cover of the Journal of the Air Pollution Control Association, July 1987 issue) and plumes being brought to the ground. Conversely these flows can enhance turbulence thus increasing the rate of dilution of the pollutant. Flow oscillations can result in periods of relative calm leading to high concentrations in a valley, or periods of intense mixing and increased wind speed where chemicals are rapidly mixed. Generally, secondary flow effects have a minimal effect on plume behavior primarily because they have scales smaller than the plume (Dickerson 1980).

5.2 MATHEMATICAL MODELLING OF FLOW FIELDS OVER RIDGES AND HILLS

Owing to the complex flow patterns in regions of complex terrain, it is of critical importance to calculate terrain effects on the horizontal and vertical flow fields. Various methods have been applied and developed.

For the horizontal flow field, the simplest approach is to assume a straight-line trajectory, i.e., the horizontal transport is unaffected by underlying terrain. This can be an acceptable assumption in situations where concentration frequency distributions over at least a year or long-term average concentrations/depositions are being calculated. On the other end of the wind field modelling spectrum are highly complex three-dimensional numerical wind field models. These models are by their very nature, large, time consuming computer programs which make them difficult to apply. Wind field model approaches can be classified as (Seinfeld 1988):

- (1) Objective analysis procedures Values of the wind field at required points are obtained by a weighted interpolation of observed data.
- (2) Diagnostic procedures Diagnostic methods use some or all of the governing equations to solve for the wind field, with the simplification that the time derivatives are all zero for the solved wind fields. These models all assume that an eventual steady-state solution adequately represents meteorological conditions for the short time period of interest, e.g., one hour.
- (3) Prognostic methods These methods are based on the numerical solution of the coupled, turbulent conservation for mass, momentum, energy and water vapor together with the appropriate thermodynamic state equations. These are the so-called "primitive equations." These methods will produce wind and temperature fields and their changes in time. They employ complex numerical schemes to solve the simultaneous differential equations.

The choice of the method depends on the particular application and the available observational data for the region of interest. A summarized description of these approaches is shown in Table 5.3.

Thomson et al. (1987) reviewed wind field models as part of the selection of a deposition model by the Western Canada Mesoscale Modelling Task Group. Other reviews are also available (Kessler et al. 1987, Seinfeld 1988, Pielke 1984). There are more than 30 such models differing in the numerical solution techniques employed and the assumptions made (especially for the prognostic models) about hydrostatic density and surface heating effects (Anderson 1971, Ludwig and Byrd 1980, Yamada 1981, Pielke et al. 1983).

Table 5.3 Summary of wind field models.

Method	Physics	Approach	Output	Advantages	Disadvantages
Objective Analysis	none	observational wind data interpolated to generate wind field	wind field at time of observations	simple, inexpensive	observations of wind in complex terrain may not be representative thus error in interpolated/extrapolated wind field interpolation/extrapolation schemes have no physical basis
Diagnostic	steady-state equations of one or more of the governing equations most common, conservation of mass	observational wind data used with mass conservation imposed and total wind-field generated iteratively	wind field at time of observation	simple, some computation required but inexpensive	error in observations create error in wind field
Prognostic	governing set of equations solved	numerical solution of equations based on observations step forward in time	wind, temperature (stability, mixing height) at future time	temperature output, better physics, forward time stepping, less spatial detail required for initial conditions	computationally expensive input observations are not necessarily conserved (i.e., reproduced)

Various wind field models have been applied to complex terrain in Alberta. These applications have been summarized in Table 5.4.

5.3 MATHEMATICAL DIFFUSION MODELLING OVER RUGGED TERRAIN

Many of the wind field models described in 5.2 are tied to a diffusion model. Some models have specific application to isolated hills or valley situations while others are for a larger-scale (mesoscale) transport where a number of hills and valleys are included. The complexity of these models range from simple modifications of the Gaussian plume model to three dimensional numerical models which can more precisely account for spatial variations in wind and turbulence fields associated with complex terrain.

5.3.1 Analytical Modelling Approaches for Ridges and Hills

The two major differences between flat terrain dispersion models and those suited for complex terrain lie in the representation of plume trajectories (horizontal and vertical) and the treatment of the turbulence field. One of the simplest approaches is to modify the plume trajectory using potential flow theory, even though there are drawbacks to applying potential flow theory to the real atmosphere.

Potential flow solutions are, strictly speaking, limited to nonstratified, inviscid flows. Thus, they can only be applied to near-neutral stability conditions and areas where there are no adverse pressure gradients which result in separation (such as in the lee of a hill). Despite the limitations, the relatively simple mathematics (Egan 1975) provides a good starting point for describing the gross features of complex flow patterns and for incorporating formulations to account for plume dispersion.

Some observational studies in Alberta confirm the consistency of airflow over an obstacle with potential flow descriptions. Others show the difficulty in using this approach. These studies are listed in Table 5.5.

The advantage of potential flow theory is its ability to provide some idea on the effect of rugged terrain on ground-level concentrations without having to resort to complex numerical computations. For illustration, potential flow streamlines are shown for flow over a hemispherical hill in Figure 5.12.

Table 5.4 Wind field models as applied to various locations in Alberta.

Reference and Model Name	Model Type and Description	Region of Application	Model Use	Model Performance	Comments
Danard & Gray (1982) BULKABL V1 and 2	Prognostic, one layer (bulk)	Athabasca oil sands, river valley	Surface winds for the region	median differences in wind speed and angle of 2.7 km/h and 19°, respectively	PC versions of model are available Galbraith and Danard (1987)
Danard and Galbraith (1987) SAM-V1	Prognostic, one layer (bulk), non-hydrostatic influences included	Mt. Allan, Canmore Valley, Canada Olympic Park	Surface winds for these regions	Comparison to six cases, wind speeds accurate to 10%, no comparison statistics for wind direction	Recommendations made for data collected to improve model performance
The Landplan Group (1984) CTWM	Diagnostic, 3-D mass continuity	Mt. Allan and surrounding region (Kananaskis)	Surface winds for the region	No comparisons to observations	Used for forecasting winds
Rudolph and Hage (1983)	Diagnostic, 2-D potential flow	North Saskatchewan River valley at Edmonton	Vortex circulation on sidewalls for use in particle-in-cell diffusion model	profiles at quarter slope and mid slope in close agreement	Wind field generated for different incident wind directions
Sakiyama (1983) WEST	Diagnostic, 3-D mass continuity	Canyon Creek and Elbow River valleys (Kananaskis)	3-D wind field for use in dispersion model	No comparisons to observations	applied to one sidewall and reflected about the valley axis to obtain complete valley winds. Wind field generated based on surface and elevated wind and temperature observations taken in valleys

Table 5.4 Wind field models as applied to various locations in Alberta. (cont'd)

Reference and Model Name	Model Type and Description	Region of Application	Model Use	Model Performance	Comments
Wong et al. (1987) UAUVM	Prognostic, 2-D specifically for valleys	North Saskatchewan River valley at Edmonton	2-D nocturnal drainage flows, temperature and wind fields generated under light regional winds and super-critical Ri #	Wind and temperature profiles as a function of time show good qualitative agreement	Drainage cell development predicted; wind field used to drive a dispersion model
Taylor (1977), Taylor et al. (1983b) Walmsley et al. (1980, 1986) MDJ3SH and variants	Prognostic, 3-D momentum equation only	Kettles Hill - Pincher Creek	3-D wind field for gently sloped, isolated hills	Array of wind and turbulence at 3 and 10 m and profiles up to 200 m Good qualitative agreement for horizontal wind field distribution Consistent overprediction of normalized wind speed above 5 m at hilltop by about 10%	Extensive field and wind tunnel data collected; see Salmon et al. (1988) for summary of project
Slawson et al. (1979) - no name	Prognostic, 1-D	Athabasca River valley at the Suncor oil sands plant	Wind and temperature profiles predicted 6 hours in advance	Qualitative comparisons with wind and temperature profiles up to 1600 m are in good agreement given observational errors	Improvements to the model are suggested (include convective penetration and baroclinicity)

Table 5.5 Observational studies in Alberta on the applicability of potential flow.

Reference	Region	Observational Study	Conclusion
Leahey and Hicklin (1973)	Crowsnest Pass near Coleman	tetroon (constant-level balloon) releases during unstable and neutral conditions; winds directed down the pass	large vertical motions due to blocking of horizontal wind on cliffs no specific assessment with potential flow, but existence of large vertical motions yield poor agreement
MEP/ Envirocon (1976)	Crowsnest Pass near Coleman	tetroon releases during unstable, neutral and stable conditions; plume photography; winds directed down the pass	neutral and stable conditions showed most tetroons moved parallel with terrain unstable conditions showed strong updrafts no specific assessment with potential flow, but existence of large vertical motions in unstable situations yield poor agreement
Leahey and Rowe (1974)	Copithorne Ridge (west of Calgary)	helicopter flights through Jumping Pound gas plant plume as it passed over ridge; neutral stability	potential flow is a conservative approximation to the observed trajectories
Chung (1979)	Copithorne Ridge (west of Calgary)	tetroon flights over the ridge under stable, night conditions (one conducted during day)	trajectories on average are in reasonable agreement with potential flow theory no systematic difference in in trajectories (for a range of natural air wavelengths from 3.0 to 1.5 km)
Slawson et al. (1978)	Athabasca River valley at the Suncor oil sands plant	plume photography when plume direction normal to the valley under neutral conditions	for west winds plume trajectory was consistent with that predicted by potential flow theory no observations for the east wind direction

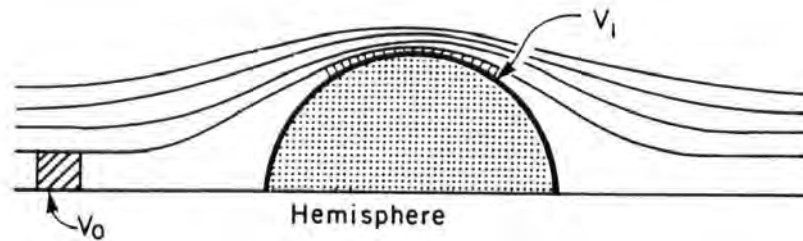


Figure 5.12 Potential flow streamlines past a hemisphere (Egan 1975).

Consider a plume segment whose vertical extent is given by the height between the surface and the lowest streamline. This segment contains a certain mass of chemical, and occupies a volume V_0 . For simplicity, assume that there is no diffusion across streamline boundaries. As the segment passes over the top of the ridge the plume gets squashed as the streamlines converge. Thus the vertical extent of the plume segment is reduced and the horizontal extent of the plume is increased. The segment also speeds up as it approaches the ridge, with its forward portion moving faster than the rear, resulting in a stretching of the plume segment. Throughout this plume segment deformation process, the volumes are maintained ($V_0 = V_1$) so the chemical concentration remains unchanged as it passed over the hill. The implication of this simplified flow model is that concentrations are completely unaffected, even though the plume centreline is closer to the ground as it passes over the hill. However, real flow over obstacles is more complex and ground-level concentrations do indeed change due to the presence of an obstacle. However, this exercise illustrates that there are compensating factors involved where the effect of an obstacle may not be as drastic as first thought.

If the role of diffusion is included, the results will be different. Wilson (1977a) outlined the changes in ground-level concentration for a ridge and an initial vertical concentration distribution given by a Gaussian shape with reflection at the surface. The ground-level concentrations more closely approach the concentration at plume centreline due to an enhancement of vertical diffusion. This occurs because of the increased vertical concentration gradients which result from streamline compression.

Wilson cautioned that the role of turbulence in the flow field may change this conclusion, since flow deformation affects the concentration gradients and the fluctuating turbulent velocities in the flow as it passes over the object.

An early attempt (Stumke 1964a, 1964b) to predict ground-level concentrations in the vicinity of terrain obstacles used potential flow theory and gradient transfer relations to arrive at an analytical solution for estimating ground-level concentrations. Because the maximum ground-level concentration is independent of source position along a streamline, a simple correction factor can be applied to the source height in flat terrain as a function of position with respect to the terrain step.

An eddy diffusivity method (Hunt and Mulhearn 1973) used in conjunction with the potential flow field over a cylindrical ridge shows an increase in the maximum concentration at ground-level ($C_{\max}$) over the flat terrain value ($C_{\max, \text{flat}}$) by the ratio:

$$C_{\max}/C_{\max, \text{flat}} = 1/[1 - h_t^2/(x^2 + h_{\text{eff}}^2)]$$

where: h_t = ridge height (semicircle height)
 h_{eff} = plume height
 x = upwind distance of the source to the ridge centre

For situations where $x \gg h_{\text{eff}}$, and $x \gg h_t$ (i.e., the ridge and plume height are small compared with the upwind distance), this equation implies that $C_{\max}$ will only be marginally larger than that associated with flat terrain.

In trying to determine the effect of a two-dimensional ridge on ground-level concentrations, Wilson (1977a, 1980) used a wind tunnel simulation to study the behavior of a nonbuoyant plume passing over a downwind hill. Three different hill shapes were considered: two bell-shaped hills of varying steepness with gradual changes in elevation, and a single triangular sharp-edged ridge with steep slopes. Various source heights ranging from ground level to twice the height of the hill crest were placed in a simulated neutral boundary layer. Based on these measurements, Wilson proposed a model which includes the effects of hills on the ground-level concentration. The assumptions in the model are that the vertical concentrations profile follows a Gaussian distribution with ground-level reflection, and that the influence of a hill is related directly to the streamline convergence or divergence experienced by the stream tube between ground-level and source streamline height. The final equation is:

$$\frac{C_{o, \text{ hill}}}{C_{o, \text{ flat}}} = \phi \exp \left(1 - \phi^2 \frac{h_{\text{eff}}^2}{2 \sigma_{zo}^2} \right)$$

$$\text{where } \phi = (1 - B_1) \frac{h_{\text{hill}}}{h_{\text{eff}}} + B_1$$

- C_o the concentration at a height h_{hill}
- h_{eff} the initial source streamline height
- h_{hill} the streamline trajectory height over the hill
- σ_{zo} vertical plume spread over flat terrain
- B_1 constant, empirically derived, Wilson suggested 0.5 on windward slopes, 0.25 on lee slopes of gentle hills that have no flow separation. Wilson also provides methods (based on different streamline trajectory descriptions) to derive this parameter for different idealized terrain shapes.

Another approach (Egan 1975) allows for a conceptual separation of the effects of kinematics (stream tube contraction) and the effects of altered turbulent diffusivities. The modifications require some estimates (either from a flow model or measurement studies) of the relative height above the surface of the plume centreline, and the streamline spacing, in the vicinity of the terrain. The modified ground-level, plume centreline Gaussian equation (including surface reflection) is:

$$C = \frac{2Q}{(2\pi \sigma_y \sigma_z u)_f (D_y D_z)} \exp \left(- \frac{(\eta h_{\text{eff}})^2}{2(\sigma_z^2)_f \zeta^2 D_z^2} \right)$$

where subscript f denotes "flat terrain" values;

$(2\pi \sigma_y \sigma_z u)_f$ = the volume flux rate of pollutant along the steam-line expressed with "flat terrain" spread statistics and horizontal velocity;

$D_y, D_z = \sigma_y / \sigma_{yf}, \sigma_z / \sigma_{zf}$, the ratios of the complex terrain vs. flat terrain spread statistics due to differences in turbulence levels;

$n = (z - h_t) / (h_{\text{eff}})$ where z is the local height of the plume, h_t is the local height of the terrain, and h_{eff} is the effective source height; and

$\zeta = (\partial \psi / \partial z)_o / (\partial \psi / \partial z)$ the ratio of the average vertical gradient of the stream function at the effective stack height to the gradient at the plume centreline over the surface.

The ζ and D_z factors are constant over plume depth. For flat terrain, D_y , D_z , η and ζ equal unity and the equation reduces to the unmodified Gaussian equation. For an infinite ridge (two-dimensional flow), $\zeta \approx \eta$ and if D_y , D_z equals unity, the deformation of the flow field has little effect on ground-level concentrations. For three-dimensional flow, because the flow may distort in the third dimension, η is not necessarily equal to ζ .

For a sphere

$$\eta \approx h_{\text{eff}}/3a$$

$$\zeta \approx 2 h_{\text{eff}}/3a$$

where a = sphere radius

If $D_y = D_z = 1$ is assumed, the exponential term in the equation becomes:

$$\exp - \frac{(0.5h_{\text{eff}})^2}{2(\sigma_z)_f}$$

For this special case, calculations are based on the assumption that the effects of the sphere on the concentrations are accounted for by dividing the effective stack height by two. This can be thought of as allowing the plume to approach the surface of the sphere to within half the effective stack height. Thus, the analytical expressions for potential flow solutions to complex flow situations provide useful insights into the effect of simple terrain features on the concentrations. However, even simpler approaches are possible, and will be outlined in the next section.

5.3.2 Adjustment Factor Approaches

The most popular ways of incorporating the effects of terrain into simple diffusion models is to lump all of the effects that a terrain feature may have on a plume into one or two correction factors that are extrapolated to a variety of situations. They are based on the Gaussian plume equation with one or two modifications to account for terrain effects and to overcome the inherent assumptions of the Gaussian plume equation which would preclude it from being used in rugged terrain areas.

To account for the plume height reduction that occurs as the plume approaches and passes over a terrain feature, a terrain adjustment factor, T_a , is used. This

factor is loosely based on potential flow theory. The basic terrain-adjusted Gaussian equation for ground-level concentrations beneath the plume centerline is of the form:

$$C = \frac{Q}{U\sigma_y\sigma_z\pi} \exp [-0.5 (T_a h_{\text{eff}}/\sigma_z)^2]$$

where T is a terrain correction factor used to adjust the effective stack height, h_{eff} .

The value of T_a may take on many functional forms. A value of $T_a=1$ means that the plume conforms perfectly to the underlying terrain. The half-height method (Hoffnagle et al. 1977) adjusts the plume height so that the plume does not approach the terrain any closer than half the plume height calculated for flat terrain. A variable height method (Lott 1986) has also been applied by Alberta Environment (1988):

$$T_a = 1 - (h_t/h_{\text{eff}}) (1-R) \quad \text{for } h_{\text{eff}} > h_t$$

and

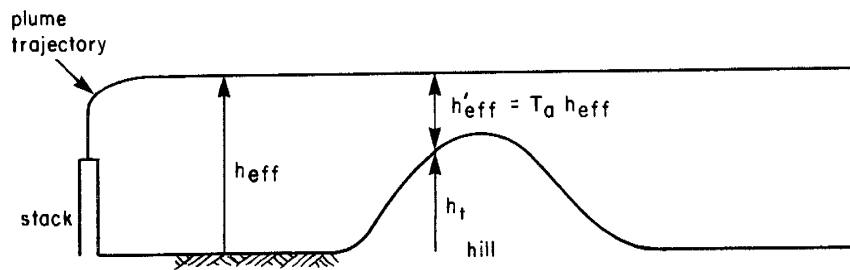
$$T_a = R \quad \text{for } h_{\text{eff}} < h_t$$

where h_t is the height of terrain above stack base, and R is a function of stability class:

Pasquill stability class R	A	B	C	D	E	F
	0.8	0.7	0.6	0.5	0.4	0.3

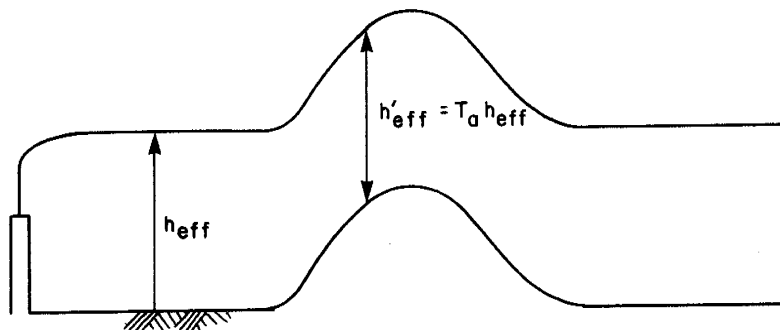
If the plume is much higher than the terrain, a simpler approach is often used. The plume remains at a constant height and the difference between h_{eff} and the elevation of the receptor above stack base is used; i.e., $T_a = (h_{\text{eff}} - h_t)/h_{\text{eff}}$, where h_t is the height of the terrain at the receptor. The mixing height is assumed to follow the terrain heights.

These "height adjustment" Gaussian-based models are commonly used for generating long-term averages or for regulatory screening purposes. They assume that the horizontal transport in complex terrain is along straight-line trajectories. The various approaches are shown in Figure 5.13. A detailed review of how rugged terrain is handled in the simple Gaussian models is given in McVehil (1988).



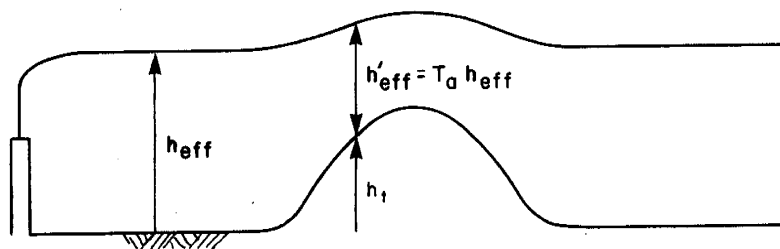
$$T_a = (h_{eff} - h_t) / h_{eff}$$

horizontal plume height assumption



$$T_a = 1$$

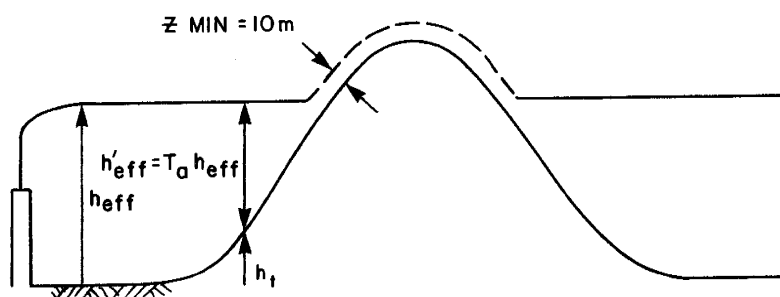
constant plume height above terrain assumption



$$T_a = (h_t / h_{eff}) (1 - R)$$

variable plume height assumption - varies with R which can be a function of stability

TERRAIN HIGHER THAN PLUME



$$T_a = (h_{eff} - h_t) / h_{eff}$$

or

$$T_a = (h_t / h_{eff}) (1 - R)$$

before point of plume impingement

at point of impingement and beyond:

$$h'_{eff} = T_a h_{eff} = Z \text{ MIN}$$

Figure 5.13 Various plume height adjustment approaches.

For some models, accounting for terrain effects has been limited to the inclusion of the T_a factor (i.e., apparently only accounting for the reduction in the plume height as the plume passes over an object). However, the effects of flow deformation, flow acceleration and changes in diffusion discussed earlier are semi-empirically incorporated in the T_a factor.

Another way to account for these other effects is to adjust the sigma values to one-half class more unstable for neutral and stable conditions (Hoffnagle et al. 1977). In effect, this increases the plume spreads for complex terrain areas over the flat-terrain values, which has some basis in observational data. Observations on plume spread in rugged terrain areas in Alberta support the increased spread approach. These are listed in Table 5.6. A review of these Gaussian-based models (EPRI, 1982) provides a list of models that account for terrain in this simple fashion.

5.3.3 Accounting for Curved Trajectories

So far the discussion has been limited to changes in the plume height with respect to the surface and plume spreads. In reality, the wind field in a rugged terrain area is not uniformly the same. That is, a plume does not travel in one direction and speed but rather changes direction and speed depending on the changes in the flow field.

There are some simple ways to approach this problem. Typically a non-uniform wind field is generated (by one of the diagnostic methods described in Section 5.2) and the plume is represented by a series of superimposed Gaussian puffs. (Allwine and Whiteman 1985, Ludwig 1984). The location of these puffs after each time step depends on the advecting wind field. In this way a curved-trajectory, Gaussian plume can be generated. The terrain correction factors described earlier can be applied to the series of superimposed puffs (Allwine and Whiteman 1985). A model similar in concept was developed for Alberta Environment by Galbraith and Danard (1987), but rather than puffs, a Gaussian plume with a variable trajectory was used in conjunction with a non-uniform wind field model, and no terrain compensation factors were applied.

5.3.4 The Special Case of Stable Situations

For plumes below the top of the terrain feature under stable conditions, the condition of plume impingement is not addressed in these simpler methods. Studies on

Table 5.6 Plume spread observations in Alberta in rugged terrain.

Reference	Location and Terrain	Observational Study	Plume Spreads Description
Leahey and Hicklin (1973)	Crowsnest Pass	tetroon releases during unstable and neutral conditions	σ_z were twice as large as flat-terrain values, depending on position in the pass
MEP/Envirocon (1976)	Crowsnest Pass	tetroon releases during stable, neutral and unstable conditions; plume photography	under very stable conditions, σ_y is similar to flat terrain, σ_z is somewhat larger
Angle and Gourlay (1983)	Elbow River valley	fast-response turbulence measurements, nocturnal drainage flow	under stable drainage flow, valley turbulence is closer to neutral, requiring a shift of 1-2 classes more unstable required
Leahey (1974)	Battle River valley	fast-response turbulence measurements and modelling work	greater turbulence than over flat topography
Leahey (1974)	Radnor Plateau	measured turbulence near plateau top for both upslope and lee flow conditions	for upslope flow, derived sigmas were from 1 to 1.5 times greater than flat sigmas; for lee flow derived sigmas were about the same to 1/2 as large as the flat-terrain sigmas
Leahey and Rowe (1974)	Copithorne Ridge	helicopter traverses of Jumping Pound gas plant plume under neutral stability	σ_z and σ_y were 1.2 and 1.7 times the respective flat terrain values

this specific situation have been extensive due to the potentially high concentrations that can result (Strimaitis et al. 1983, Lavery 1983, Strimaitis et al. 1985, and Dicristofaro 1985). High concentrations on hillsides can occur due to plume impingement or streamline depression where the plume is pushed closer to the surface than would occur in flat terrain.

In order to determine whether a plume will impinge on a hill, the concept of a *dividing streamline* was developed representing a division between two regions of flow that occur with stable flow around a terrain obstacle (Sheppard 1956, Drazin 1961, Kao 1965, Hunt et al. 1979 and Snyder et al. 1985). Rowe (1980) shows mathematically that below the dividing streamline, the flow has insufficient energy to surmount the terrain feature and consequently passes around it in a horizontal plane. Above the dividing streamline, the flow passes up and over the terrain feature. A stagnation streamline exists below the dividing streamline but not above it. This situation is shown in Figure 5.14.

The concept of the dividing streamline can be illustrated and predicted using the *Froude number*

$$Fr = \frac{U}{h_t \left(\frac{g}{T} \frac{\partial \phi}{\partial z} \right)}$$

where:	T	temperature
	g	acceleration due to gravity
	h _t	height to top of terrain
	$\frac{\partial \phi}{\partial z}$	potential temperature gradient
	U	wind speed

The Froude number can be interpreted as a ratio of inertial to buoyancy forces in a fluid. The height of the dividing streamline, h_d, is given by

$$h_d = h_t (1 - Fr)$$

This equation assumes a uniform upstream velocity profile and a consistent density gradient.

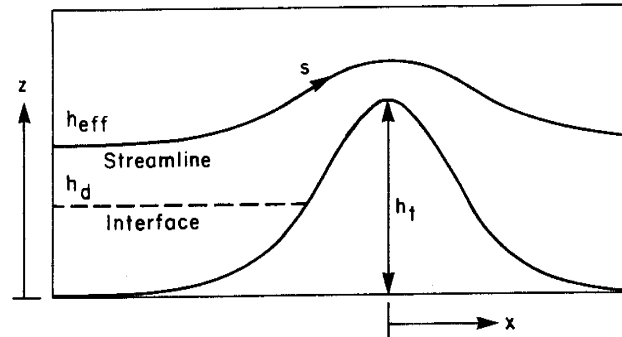


Figure 5.14 A schematic diagram of a stable flow streamline across a terrain obstacle (Rowe 1980).

h_{eff} = upstream elevation of streamline

h_d = elevation of dividing streamline between the two layers

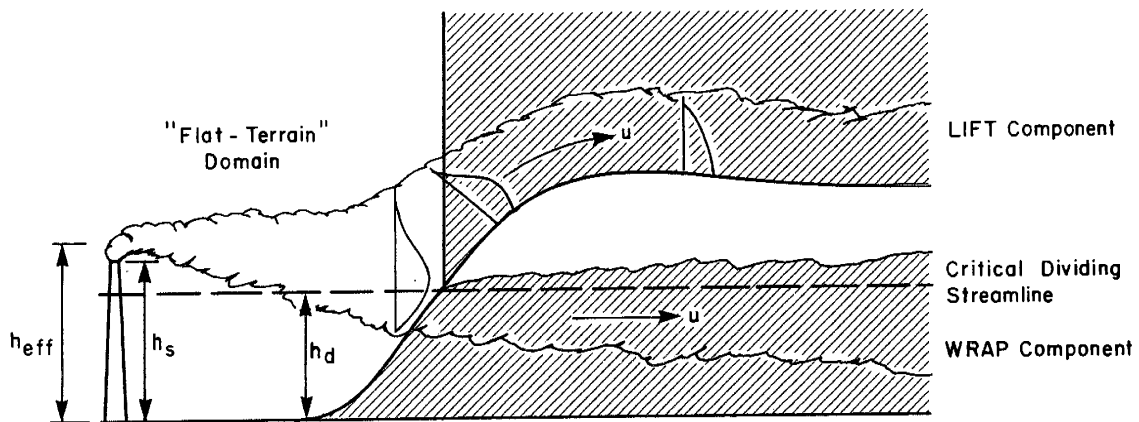


Figure 5.15 Idealized stratified flow about hills (Venkatram 1986).

The dividing streamline provides for a conceptual framework called the two-layer or split plume method for modelling in such situations. The split plume method allows for the portion of the plume above the dividing streamline to go over the hill, and portions of the plume below this point to go around it (this is illustrated in the LIFT and WRAP components of the plume shown in Figure 5.15). The EPA model CTDM (Strimaitis et al. 1987) uses this framework for concentration estimates when a plume interacts with a hill in stable flow.

Evidence supporting the two-layer approach in Alberta has been found from observations conducted at Copithorne Ridge, southwest of Calgary. Benjamin et al. (1977) conducted a flow visualization experiment over this ridge during the summer to examine the effect of stable flow towards the ridge. Smoke trails up to 200 m above ground were generated and a photographic method was used to examine the smoke trajectories at various heights above the terrain. Observations showed that the ridge induced horizontal (directional) shear. This indicated that plume impingement could occur below some dividing streamline height and explained the high frequency of peak concentrations observed under these conditions.

Time lapse photographs of the gas plant cooling tower plume as it passed over Copithorne Ridge were taken during two winters and reported by Lee (1979) and Rowe et al. (1982). The plume rose to a height slightly greater than the ridge itself, thus the influence of the ridge on the plume behavior can be clearly seen. Four cases were identified as suitable for generating trajectories and comparing them to single-layer streamline trajectories and the trajectories from a two-layer flow model. A typical trajectory is shown in Figure 5.16. Modelling the flow based on a two-layer approach fits the observations much better than the single-layer potential flow.

The results provided the basis for a simple screening model proposed by Rowe and Tas (1986) for applications to stack design in complex terrain. The critical design condition for a stack is on the effective stack height such that the plume just reaches the topmost layer of the most stable layers likely to occur. They developed a simple approach that identifies the case where the plume rise above stack height equals interface height above stack height. By nondimensionalizing these two heights a simple nomogram was developed (Figure 5.17). The curved line is the nondimensional plume rise function and the straight lines are the nondimensional interface heights for different

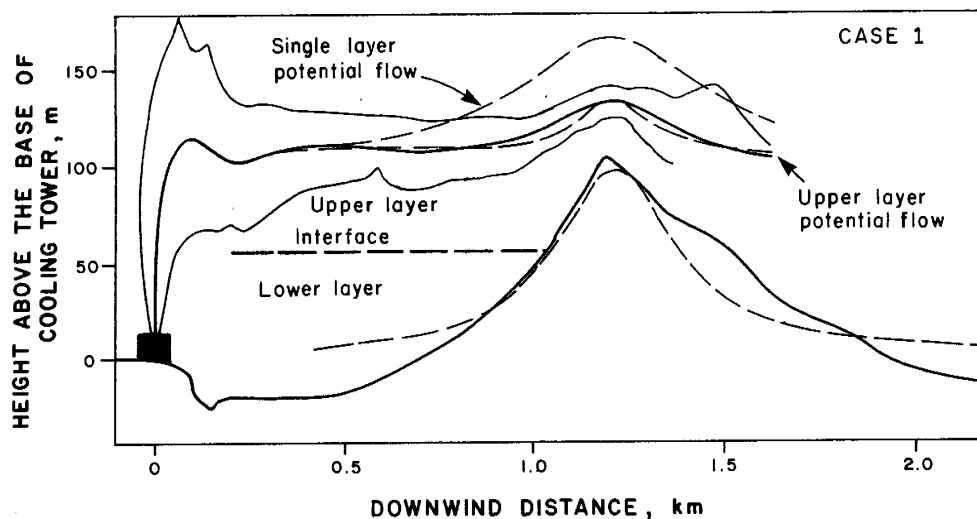


Figure 5.16 Cooling tower plume outline and centreline trajectory. The potential flow streamlines for both the single-layer and the two-layer models are also shown. The dashed curve next to the ridge denotes the theoretical bell-shaped ridge (Rowe et al. 1982).

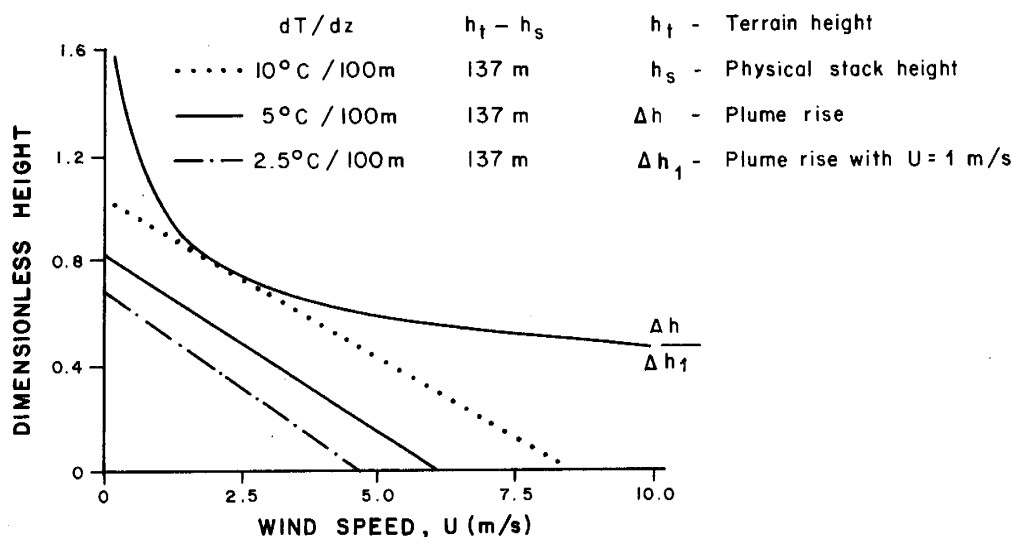


Figure 5.17 Dimensionless elevation (plume rise or interface elevation above stack top) as a function of wind speed (Rowe and Tas 1986).

temperature gradients. The point of intersection between these two curves occurs at the critical wind speed. For wind speeds higher or lower, the plume will be above the dividing streamline. Rowe and Tas showed that:

- (1) if the effective stack height is made large enough so that the plume centreline just reaches the upper layer flow in the most stable atmosphere that is likely to occur, the plume would be lifted clear of the terrain in a less stable atmosphere;
- (2) if it is required that the plume should always be placed well into the upper layer flow, then the plume rise at 1 m/s in the most stable atmosphere should be 1.5 times the height of the terrain crest above stack top.

5.3.5 Numerical Models

The analytic methods described so far are very popular because of their relative simplicity and ease of use. Gaussian puff and plume models continue to be used with success even when pushed beyond their theoretical limits. However, analytical methods are typically inflexible under complex atmospheric conditions: i.e., variable winds and thermal stratification over irregular surfaces. In addition, in areas of complicated underlying terrain (combinations of hills and ridges) complex flow patterns due to blocking and channelling result in trajectories which cannot be adequately described by simple potential flow approaches. The tetraon trajectories reported by Leahey and Hicklin (1973) and MEP/Envirocon (1976) in the Crowsnest Pass (see Table 5.8) showed the difficulties in applying simple (analytic) potential flow models to these situations.

An alternative seems possible only with the coupling of one of the complex wind-field models and a sophisticated numerical diffusion model. These models also have problems, since numerical damping/diffusion and parameterizing turbulence (via K theory) can result in unrealistic predictions. A first-order numerical diffusion model solves the advection/diffusion equation for a fixed grid system:

$$\partial C / \partial t + \nabla U C = \nabla (K \nabla C) + S + R$$

where

C	- pollutant concentration
U	- 3-D wind field
K	- Fickian diffusiveness
S	- source term
R	- formation/destruction due to chemical reaction
∇	- gradient operator ($\partial / \partial_x, \partial / \partial_y, \partial / \partial_z$)

Three-dimensional *grid models* typically (Fabrick et al. 1977, Tesche et al. 1987) solve this equation by a finite differencing method. There is also considerable interest in applying finite element methods since they offer many advantages over differencing methods. However, solutions of large three-dimensional problems require large computer resources. A more fundamental form of this equation can be solved where the turbulent transport terms are represented by second-order closure terms. These *second-order models* are by their very nature complex and computer intensive (Yamada 1979, Yamada and Bunker 1988).

Particle-in-cell, three-dimensional models are also popular alternatives used in complex terrain models (Lange 1978). Various names have been given to these models such as particle-in-cell, Lagrangian statistical-trajectory and Lagrangian-Eulerian. A typical example is the MODTRAC model developed by Davison et al. (1985). The plume is represented by thousands of fluid elements (particles) tracked individually through the atmosphere. Each element is subject to advective transport by a prescribed mean wind field, and to turbulent motion by random velocity fluctuations, which can be generated by a Markov chain relation given certain constraints on model time steps. The advantages of this technique over more traditional approaches to dispersion modelling are: reduction of numerical diffusion errors and improved flexibility in simulating particle deposition and linear transformation processes (Wilson et al. 1981, and Sawford 1982). The flow and turbulence fields are usually defined by a diagnostic model, then the particles are released into this field and tracked. Reid et al. (1979) concluded that a model of this type (ADPIC) would be useful for the oil sands region in northern Alberta. However, the computer resources required to run such a model make it an expensive modelling technique.

Table 5.7 is a summary of numerical models that have been applied to complex terrain areas in Alberta. Thompson et al. (1987) and Western Mesoscale Modelling Task Group (1985) provide a wide review of various numerical models.

5.4 THE EFFECT OF BUILDINGS ON PLUME DISPERSION

Buildings can be thought of as terrain objects which can substantially alter the transport and diffusion of pollutants. The flow fields vary widely with building geometry and wind directions (Strom 1976). Reproducible results in field experiments have been almost impossible to obtain (Munn 1966) and most knowledge about such disturbed airflows comes from wind tunnel studies (Sherlock and Leshner 1954; Strom 1955, 1956; Strom et al. 1957; Halitsky 1962, 1968; Hosker 1984). Even though the flow and diffusion features tend to be unique to the particular building/release characteristics, some general behaviors have been identified.

5.4.1 Shear Flow Near Buildings with a Ground Plane

The principal characteristics of the flow field around a building and plume interaction are shown in Figure 5.18. The main feature is an aerodynamic distortion zone around the building which consists of three zones -- the displacement zone, the cavity and the far wake (Halitsky 1968). The *displacement zone* is an area of displaced flow around the building. The *cavity* is the quasi-steady recirculating flow zone immediately downwind of the building. The *far wake* is the zone of reduced wind speeds and enhanced turbulence downwind of the cavity. This conceptual model of the flow around the building has lead to simple stack design rules of thumb that will be discussed later.

Recent investigations of flow in the vicinity of buildings reveal many complexities (Figure 5.19). The general characteristics of this more complex flow pattern are as follows:

- (1) There is a standing vortex in the front of the building with downward flow at the building face and upwind flow away from the building at ground-level. The size and strength of this vortex depends on the incident wind profile and the building geometry.

Table 5.7 Numerical model applications for complex terrain areas in Alberta.

Model Name and Reference	Model Description	Region of Application	Comments
INTERA/EMS Lanz (1972)	Eulerian, 3-D grid solution, flow field generated by modified potential flow	Athabasca oil sands region, Intercomp (no date)	Suncor plume behavior studied. Most severe air quality conditions were during calms where SO ₂ pooling occurred
		Balzac gas plant, Intercomp (1972)	Gas plant behavior under fumigation conditions studied.
ITSSAPP Wallis et al. (1975)	Eulerian, 2-D grid solution, flow field generated by potential flow	Battle River power plant, Wallis et al. (1975)	Maximum concentrations resulting from power plant plume, model predicted to within 6% of observed maximum
		Wildcat Hills gas plant, Wallis et al. (1975)	Ground-level concentrations higher and closer to the source than what is predicted by flat terrain model.
IMPACT Fabrick et al. (1977)	Eulerian, 3-D grid solution, diagnostic wind field model	Elbow River/Canyon Creek valleys, Sakiyama and Angle (1988)	NUVAL (a valley version of IMPACT) used for nocturnal drainage situations, relative mean absolute errors of 24-59% achieved when compared to tracer gas data observations
no name - Rudolph (1980)	Eulerian, 2-D particle in cell model, wind field given as valley cell circulation	North Saskatchewan River at Edmonton, Rudolph (1980)	Simulations of pollutant transport in valley circulation cell, initial source diffusion important to predictions, return flow important in pollutant distribution
no name - Wong (1985)	Eulerian, 2-D particle in cell model, wind field given by prognostic scheme of Wong et al., (1987)	North Saskatchewan River valley at Edmonton, Wong (1985)	CO concentrations predicted high impacts at the valley top and bottom convergence zones, observed increases in concentration simulated only when no prevailing wind assumed

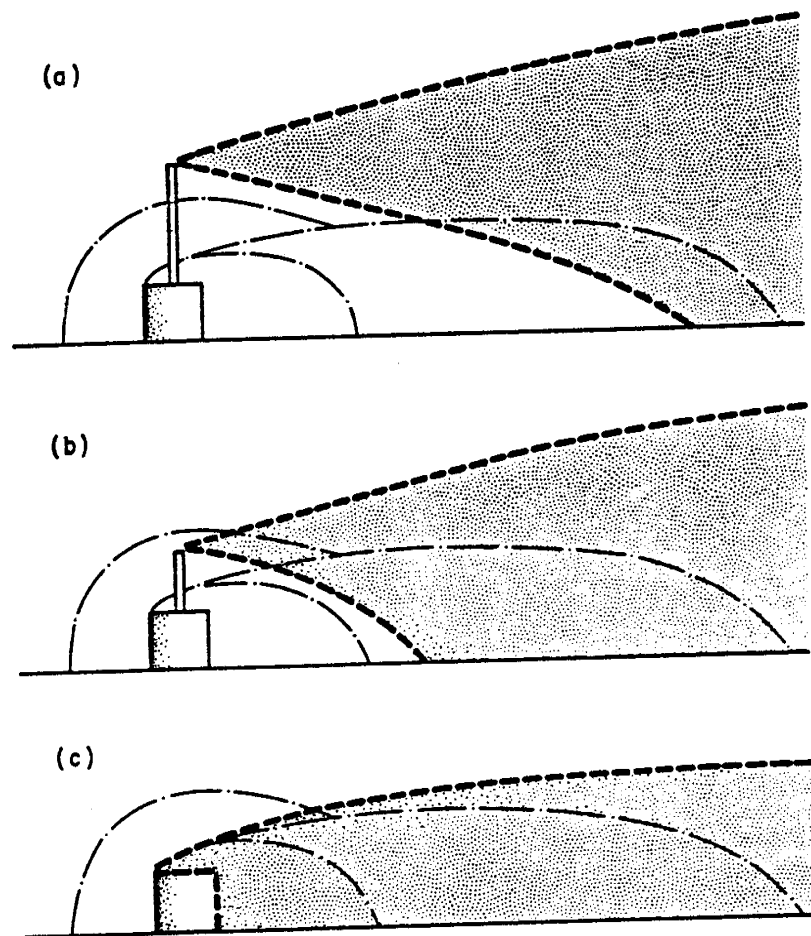


Figure 5.18 Behavior of plumes emitted directly above a building (Smith 1968).

- (a) Reduced wake intersection.
- (b) Plume enters displacement zone but not cavity.
- (c) Cavity is filled with uniform high concentration.

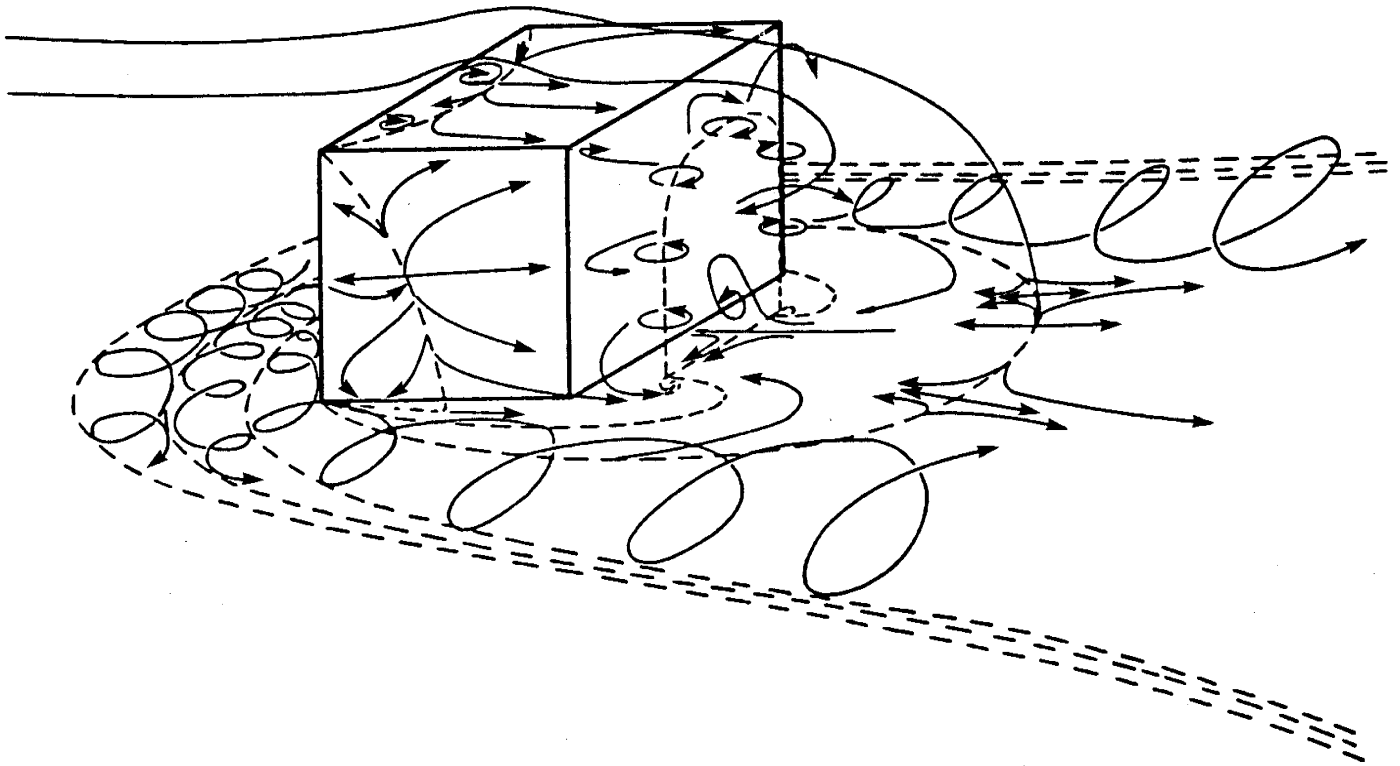


Figure 5.19 Flow pattern around a rectangular block with re-attachment of the free shear layer (Woo et al. 1977).

- (2) The interaction of the incident flow with the standing vortex results in a vortex that is wrapped around the building (commonly referred to as the horseshoe vortex).
- (3) At the leading edges of the building, the flow cannot turn sharply enough to follow the roof and sides, and therefore separates from the building. A recirculating flow is found in these zones of separation. Building geometry can dictate the extent of the separation zone.
- (4) The flow patterns behind the building consist of the continuance of the horseshoe vortex around and behind the building. In the lee of the building there is a pair of vertical vortices that interacts with the main flow near roof level and trails off downwind as an elevated counter-rotating pair. The size of the cavity zone is strongly influenced by building geometry and where re-attachment of the separation zones occurs.
- (5) The wake (downstream of the recirculating cavity zone) is characterized by turbulence enhancement and a velocity excess or defect. The velocity excess will occur if organized vorticity is present in the wake (for block buildings, this occurs when the incident flow is not perpendicular to the building face). No organized vorticity in the wake is present when the ambient flow is directly at the building face. The rate of decay of the excess or defect is a function of downwind distance and building height.

The presence of varying levels of turbulence in the ambient incident flow results in changes in the characteristics of the flow near and downwind of a building. Turbulence changes the wake cavity size and shape, weakens the vortex formation process and enhance mixing and entrainment in the separated shear layers that bound the cavity (Hosker 1984).

5.4.2 Clustering Effects

Most studies of the flow near a cluster of objects have dealt with specific industrial or nuclear complexes or with urban street canyons. Thus general principles

are sometimes difficult to elucidate (Hosker 1984). However, different combinations of building clusters show recirculating vortexes and venturi effects. These are shown in Figure 5.20. Urban air quality studies (Cermak et al. 1974, Hoydysh et al. 1974 and Wedding et al. 1977) suggest that it might be useful to scatter high-rise buildings throughout a city to reduce concentrations. Chemicals are moved upward by vertically oriented spiral vortexes in the wake of high rise buildings while cleaner air aloft is diverted downward along the windward face of the building. The result is a reduction of chemical concentrations near ground level. In addition, placement of the urban array at 45° to the incident wind produced the best overall urban air quality.

5.4.3 Isolated Sources Upwind of Buildings

In the situation where a plume is impinging on a building, a number of different behavior patterns can occur. The incident flow near the top third of the building is directed up, taking the plume with it, while the downwash generated by the frontal eddy in the lower portion of the building face mixes the plume downward. This results in high ground-level concentrations. Some material is carried off in the trailing horseshoe vortex. Chemicals in the cavity are well mixed. If the along-wind length of the building is large, the flow which has separated from the upwind roof edge may re-attach and bring the pollutant closer to roof level resulting in high concentrations. If there is no re-attachment, the incident plume is deflected above roof level.

In general, the overall effect of the building is to induce a rapid enlargement and dilution of the incident plume (Hosker 1984). Ground-level concentrations near the building are higher than would be observed if the building did not exist; however, further downwind, concentrations are lower than would be observed if the building did not exist.

The behavior of a plume released at roof level when it impinges on a downwind building was studied by Wilson and Netterville (1978) in the University of Alberta wind tunnel. They found that building downwash of the plume causes ground-level concentrations in front of and immediately behind the building to increase by factors of 2 to 5. They suggest that for the purposes of stack design, the concentration on all building surfaces is uniform and equal to the average between the rooftop and ground-level concentrations computed for flow in the absence of buildings. In addition, a conservative estimate for a concentration at ground-level would be to take the maximum concentration intercepted at any point on the building surface.

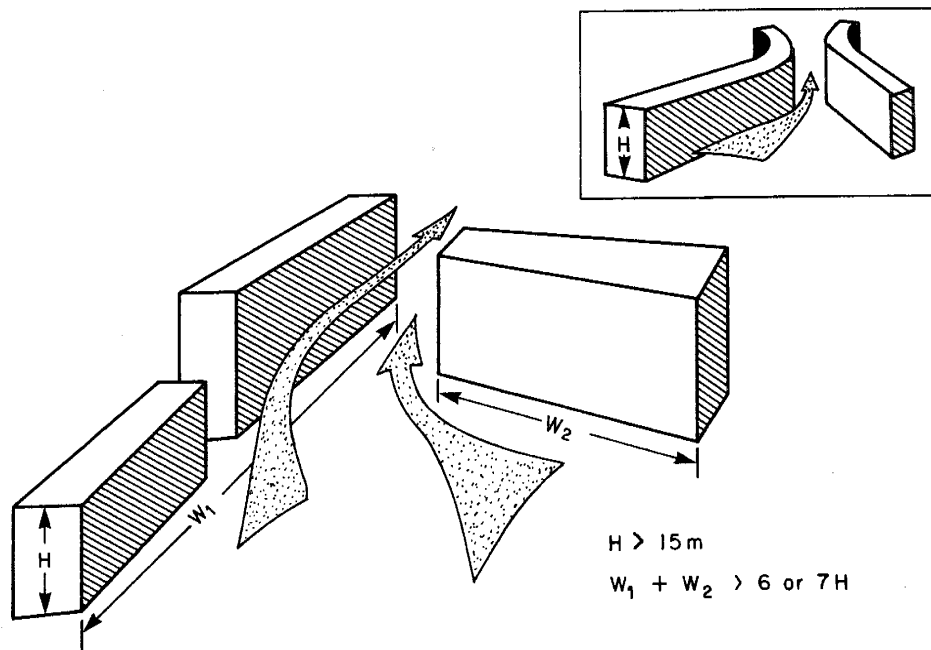
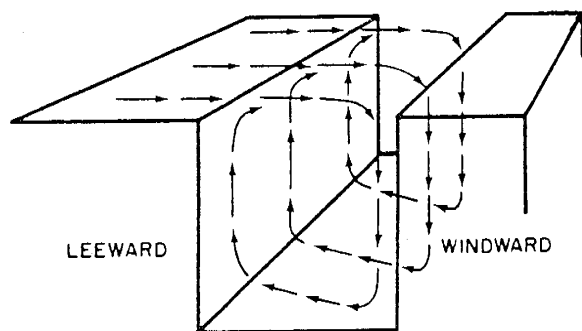
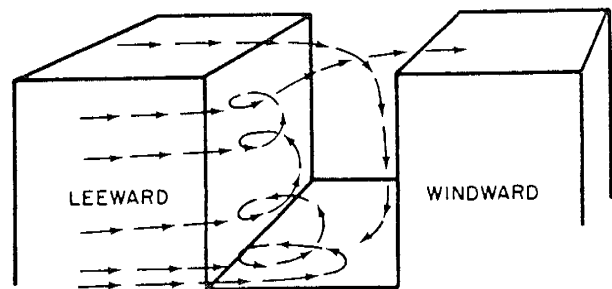


Figure 5.20a Venturi effect between buildings (Gandemer 1976).



ROOF VORTEX CIRCULATION

Figure 5.20b Recirculation in a two-dimensional street canyon (Hoydysh et al. 1974).



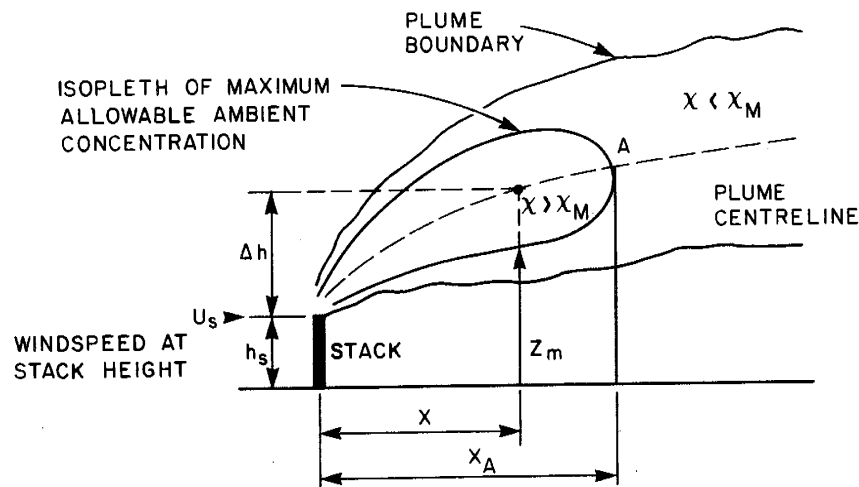
ROOF VORTEX PLUS CORNER VORTEX

Figure .20c Three-dimensional interaction of corner and cavity vortex near edge of a city block (Hoydysh et al. 1974).

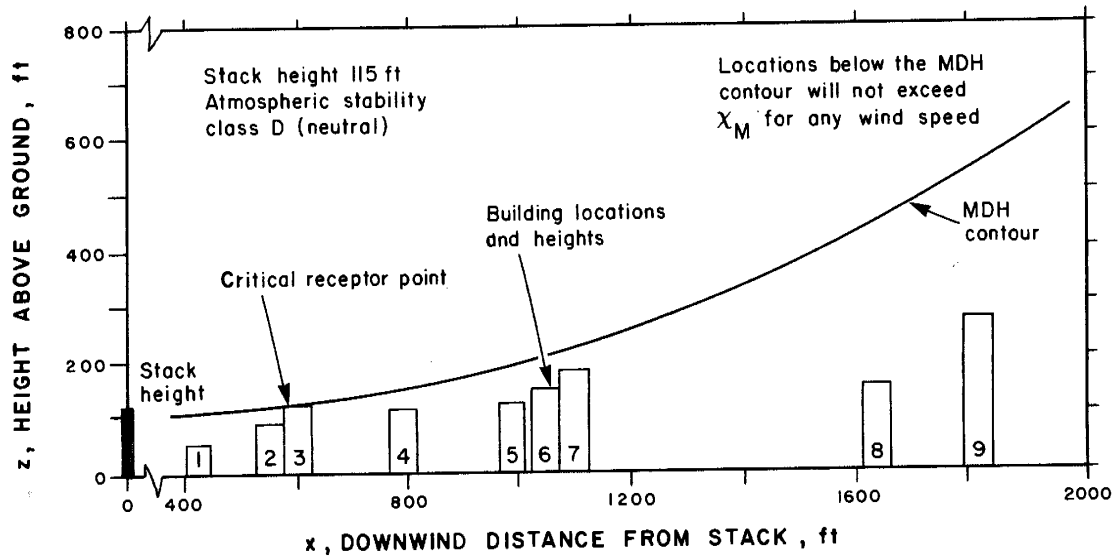
Wilson and Netterville (1976) provided a design procedure which determines the required stack height such that any given critical concentration is not exceeded in a building downwind of a source, since windows or air intake vents (which are commonly located near the tops of buildings) will allow chemicals to enter the building. The technique is based on their finding (from wind tunnel studies) that when a plume impinges on a building, the result is a fairly uniform concentration between roof level and ground. Thus, for design purposes, the concentration at roof level should be kept below allowable limits regardless of the location of building intake vents. The approach uses the Gaussian plume formulation and an iteration routine to determine a minimum descent height (MDH) given a wide range of stability and wind speeds. At a given downwind distance and stability, concentrations vary as a function of wind speed. In addition, the height above ground of a particular concentration isopleth is also a function of wind speed. The MDH is the minimum height that the specified critical isopleth will attain for a range of stabilities and wind speeds. The top of a downwind building must be below the MDH in order to ensure that building concentrations will not exceed the critical values.

The technique was applied to the design of the University of Alberta's powerhouse stack, located near several high-rise buildings. Figure 5.21 shows the relationship of the MDH contour and the building heights. Building 3 proved to be the critical receptor point and the stack had to be redesigned so that the building height did not exceed the MDH. A simpler method to generate the MDH line is based on the fact that the downwind extent of the maximum concentration isopleth changes with wind speed. At the point where the downwind extent is a maximum, the height Z_m shown in Figure 5.21 is set to Z_A . The simplification assumes that Z_A equals the effective stack height. This alternate simple approach is not conservative and the errors can be in the order of 20% near the source. However, the errors are less than 2% for downwind distance to stack height ratios of 20 or greater.

Wilson and Netterville (1976) make two important points in the consideration of this situation. First, they used rural σ values since they supply a safety factor over the use of the larger σ values characteristic of built-up areas. Secondly, in situations where there are no buildings, critical atmospheric conditions may be unstable or trapping/fumigation situations whereas the MDH curve is closest to the ground under neutral conditions. This emphasizes the need to examine the whole range of plume behavior.



(a)



(b)

Figure 5.21 Definition (a) and sample calculation result (b) for the minimum descent height (MDH), Z_m , of contour of maximum allowable concentration (Wilson and Netterville 1976).

5.4.4 Sources Near Buildings

When a stack is located near a building, the resulting effects on the transport and diffusion are complex. Vet et al. (1981) summarized the possible approaches to modelling the concentrations downwind of a gas compressor station.

One simple approach is the "2.5 times the building height" rule: if the stack height is greater than 2.5 times the building height, the plume is unaffected by the presence of the building. Otherwise a plume released within the displacement zone would be caught in the far wake, bringing the chemicals to the ground (with accompanying high ground-level concentrations) earlier than if there were no building. Even worse, the plume could end up in the cavity zone, where recirculation and continued accumulation results in very high concentrations not only in the cavity, but also downwind as turbulent transfer from the cavity to the far wake occurs. This rule of thumb was originally proposed by an English committee in 1932 (Hawkins and Nonhebel 1955, Committee 1932).

For a stack height of 1.5 building heights, the plume enters the displacement zone but not the cavity (Halitsky 1968). Even though this is not an ideal situation for reasons outlined above, the very adverse situation of cavity entrapment is avoided without resorting to very high stack heights which may be indicated by the 2.5 rule. The 1.5 rule was proposed by Angle et al. (1987) as a minimum stack height for compressor facilities, and was adopted by the ERCB and Alberta Environment (1988).

Alberta Environment (1980) recommended a generalized approach based on two length scales (Briggs 1975, Snyder and Lawson 1976). The method assumes that a plume will escape the building influence completely if the stack height, h_s is given by $h_s = h_b + 1.5 L$ (where h_b is the building height and L the smaller of either h_b or the maximum building width perpendicular to the wind direction). For a squat or cubical building this is equivalent to the 2.5 rule, but relaxes the rule for a tall, slender one.

If the height given by this rule is far in excess of what would be needed to meet the ambient air quality objectives, this criterion can be eased somewhat. The approach is to assume that the effect of the building can be accounted for by lowering the actual stack height to a value of h'_s . Thus in all subsequent calculations of ground-level concentrations, h'_s should be used instead of the actual stack height, h_s , where

$$h'_s = 2 h_s - (h_b + 1.5 L)$$

In no case may a stack ever be less than:

$$h_{s\min} = h_b + 0.5 L$$

which is the 1.5 rule for $L = h_b$

If the original stack height is equal to $h_b + 1.5 L$ then $h'_s = h_s$ (i.e., the corrected stack height = the original stack height). This simply states that the building, with this stack height, has no effect on the plume and thus no correction is required.

These simple approaches have now given way to more complete and explicit methods which can account for building effects without resorting exclusively to stack height considerations. Vet et al. (1981) reviewed 10 modelling approaches to the problem of building effects on the downwind concentration and determined four suitable candidates. With further evaluation, which included basic physical modelling principles and performance of individual models with respect to other models and experimental data, the authors recommended either the Huber-Snyder or the dual plume models as useful for compressor station installation stack design.

The Dual Plume model (Engineering Science 1977, 1980) has the following characteristics as listed by Vet et al. (1982):

- (1) A cavity entrainment factor which determines the portion of time that the plume is considered to be either elevated or at ground-level.
- (2) The portion of the plume entrapped by the cavity region can be determined for any combination of wind speed, effluent velocity, building and release height.
- (3) The portion of the plume free from the cavity behaves as if no building were present and disperses as a Gaussian plume.
- (4) The portion of the effluent captured by the cavity is subject to initial mixing in the cavity and then disperses downwind as a Gaussian half plume.

- (5) The ground-level concentration downwind of the building is the sum of the concentrations resulting from the two portions.

The downwind centreline concentrations are determined by the following equation:

$$\chi = \frac{1}{U} \left\{ \frac{aQ}{\pi\sigma_y\sigma_z + CA} + \frac{(1-a)Q}{\pi\sigma_y\sigma_z} \exp(-h^2/2\sigma_z^2) \right\}$$

where A is the projected cross-sectional area of the building facing the wind, a is the mean fraction of the plume entrapped within the building cavity, and C is a dilution factor related to cavity area, generally a value between 0.5 and 2.0.

The first term in the equation models the dispersion of that portion of the plume which is entrapped in the cavity. The second term models the dispersion of the plume which is unaffected by the building.

In the next study phase Vet et al. (1982) revisited the review of the models and chose the dual plume model as the basis for further development of a model with applicability to compressor station facilities (the new model was subsequently called CSCM). The split plume design made it possible to account realistically for the extra NO_x to NO reaction time experienced by the portion of the plume trapped in the cavity. Comparisons between monitoring data collected at the Pendant D'Oreille compressor station and CSCM predictions showed that CSCM predicted much smaller concentrations than were being observed. These differences were attributed to other buildings influencing plume behavior (CSCM only accounts for the effects of a singular building whereas the site had three buildings) or incorrect stack exit velocity resulting in wrong plume rise estimates and inappropriate entrainment factors. The comparison was not a rigorous test of model performance.

The Huber-Snyder model (Huber 1977, 1984; Engineering Science 1980) accounts for building wake effects using enhanced vertical and sometimes horizontal dispersion coefficients. The cavity entrainment factor, a key to the dual plume model, is not used here. The enhanced dispersion coefficients are based on the standard sigma curves modified for building wake effects as a function of downwind distance and building geometry. This approach is being used in regulatory models in Alberta (Sakiyama et al. 1988). The wake-affected sigmas are as follows:

$$\begin{aligned}
 \sigma_y' &= 0.7 HW/2 + 0.067 (x-3HB); \text{ for } 3HB < x < 10HB \\
 \sigma_z' &= 0.7 HB + 0.067 (x-3HB); \text{ for } 3HB < x < 10HB \\
 \sigma_y' &= \sigma_y(x + S_y); \text{ for } x \geq 10 HB \\
 \sigma_z' &= \sigma_z(x + S_z); \text{ for } x \geq 10 HB
 \end{aligned}$$

where:

x = distance downwind from the leeward edge of the influential building.

HW = width of the region of building influence normal to the wind.

HB = height of the influential building.

σ_y
 σ_z = dispersion parameters without the influence of influence of buildings.

S_y = the virtual source distance such that $\sigma_y' (10HB) = 0.7 HB^2 + 0.5 HB$.

S_z = the virtual source distance such that $\sigma_z' (10HB) = 1.2 HB$.

For a building that is tall (this has been defined here as $HB > HW$) HW should replace HB in estimates of σ_z' and σ_y' . For very squat buildings ($HW > 5HB$), HB should replace HW in the σ_y' estimate.

If the stack height is greater than 2.5 HB , the plume is assumed to be unaffected by building downwash. If the stack height is greater than 1.2 HB , only σ_z' is affected.

Sakiyama et al. (1988) used NO_x/NO_2 monitoring data taken at six compressor station installations in Alberta and did a simple comparison study with a model (named SEEC) using the Huber-Snyder approach. For data *unpaired in space and time*, the maximum NO_x and NO_2 and the 10 highest concentrations were compared. Average bias, relative mean obsolete error, dimensionless mean square error and the linear correlation coefficient were used to determine model performance. The highest NO_x and NO_2 were predicted quite well; however, performance for NO_2 was not as good as for NO_x . This indicates a weakness in the conversion chemistry module.

The Huber models I and II (Huber 1977) were used to evaluate the downdraft effects from the Keephills plant at Keephills, Alberta. The description and results of the combined wind tunnel modelling study are reported in Leahey et al. (1979) and Western Research and Development (1978). The Huber model I is the same as that used in SEEC

and the model II is a simpler approach where:

$$\begin{aligned}\sigma'_y &= [\sigma_y^2 + (.35 HW)^2]^{1/2} \text{ for } x > 3HB \\ \sigma'_z &= [\sigma_z^2 + (.35 HB)^2]^{1/2} \text{ for } x > 3HB\end{aligned}$$

HB should replace HW in estimates for σ'_y for buildings that are much wider than they are tall. The study showed that the model II almost matched the wind tunnel results, whereas the model I approach yielded overpredictions. The results indicated that a stack 1.75 times the building height at Keepphills would be sufficiently tall to ensure acceptable air quality.

5.5 PHYSICAL MODELLING OF TRANSPORT AND DIFFUSION OVER COMPLEX TERRAIN

Throughout this chapter there have been references to wind tunnel modelling studies where a scale model is placed in a wind tunnel and various flow features are observed. This, in contrast to mathematical modelling, is called *physical modelling* and it is a popular approach to study various aspects of transport and diffusion in complex terrain (Environment Ontario 1987) as well as a host of other processes such as air-surface interchanges and exchanges (Raupach and Thom 1981), urban pollutant transport (Wedding et al. 1977, Cermak et al. 1974), effluent density effects (Meroney and Neff 1979), and seashore/land flow interactions (Ogawa et al. 1975). Physical modelling is not necessarily limited to either wind tunnels or water channels, although these are the most commonly used devices. Other devices include environmental chambers, smog chambers and electrolytic tanks (Hosker 1982).

The primary advantages of a physical modelling studies for investigating the transport and diffusion of pollutants in areas of complex terrain are:

- (1) easier, cheaper, and more feasible than full-scale,
- (2) ability to modify the physical model to look at future scenarios (new buildings, large excavations) and their effects on the flow field,
- (3) ability to examine a variety of easily reproducible atmospheric conditions and the resulting flow field,

- (4) events in a wind tunnel occur 20 to 30 times more rapidly than their full-scale counterparts, allowing larger time averages to be generated in a relatively short time (Wilson 1979a).

Physical modelling can also be thought of as a unique type of analog computer (Snyder 1972) with "near-infinite resolution" and "near-infinite memory" which employs real fluids and not mathematical descriptions of their behavior.

There are, however, difficulties with the use of physical modelling. Apart from the problems associated with achieving appropriate geometric, kinematic, dynamic and thermal scaling, physical modelling is limited to time and length scales which include microscale and some mesoscale phenomena (Avissar et al. 1990). In other words, physical modelling is limited to the small-scale motions that are associated with roughness elements, small topographical features and weak differential surface heating in the boundary layer. In addition, where plume buoyancy effects are important (stable or unstable boundary layers) a host of other problems become important such as hardware and scaling limitations. Unfortunately the type of flow that has the best chance of being simulated well, the neutral boundary layer, occurs only a fraction of the time in the real atmosphere. Thus, results from a physical modelling study, when applied to the real atmosphere, must always be viewed with caution. Confirmation of physical modelling results with actual field observations is needed, as in the fluid model simulations and corresponding field observations of the flow and turbulence over Kettles Hill, Alberta described by Salmon et al. (1988); the strong convection study in Colorado (Moninger et al. 1983) and the Rakaia Gorge study (Meroney et al. 1978) in New Zealand.

5.5.1 Fundamental Principles

In order for the physical model tests to represent the same behavior as the full-scale situation, general equivalence between the model and full scale must be duplicated. The general similarity requirements are established by dimensional analysis.

Dimensional analysis yields dimensionless parameters such as Rossby, Reynolds, bulk Richardson, Eckert, Euler, Prandtl, Schmidt and Peclet numbers. These parameters should be the same for the model and prototype if the two flow systems

are to be exactly the same (Avissar et al. 1990). Equality of these numbers must be supplemented by the requirement that the surface boundary conditions and approach flow characteristics be similar for the atmosphere and the physical model. Boundary condition similarity requires similar values of:

- (1) surface roughness,
- (2) topographic relief,
- (3) surface temperature distribution,
- (4) upstream distribution of mean and turbulent velocities,
- (5) mean and turbulent temperatures, including inversion height, and
- (6) longitudinal pressure gradient.

Invariably, because of the small scale of the models, matching all of these nondimensional parameters is impractical or impossible. Hence the modeller must decide on the basis of the particular application, which parameters are critical in terms of matching and which ones are not. An example of this is the neglect of Coriolis forces on the flow phenomena studied, where the nondimensional Rossby numbers for the model and full scale will not be matched. This mismatch is acceptable for small-scale flows where Coriolis forces can be neglected (i.e., neglecting the turning of wind direction with height).

Another example of mismatching is the Reynolds number criterion. The scale reductions commonly used in physical models result in model Reynolds numbers 3 to 4 orders of magnitude smaller than found in the atmosphere. If strict adherence to the Reynolds number criterion were required, no atmospheric flows could be modelled. Various arguments have been presented to justify the neglect of the Reynolds number criterion in terms of laminar flow analogy, Reynolds number independence, and dissipation scaling. In essence, what these analogies mean is that the gross features of turbulence are similar over a wide range of Reynolds numbers (Townsend 1956).

Generating the appropriate boundary conditions is also a critical factor in the proper application of physical modelling. Central to this issue is the equivalency of the turbulence in the model to that of the real atmosphere. For the study of plume dispersal, not only must the turbulence intensities be properly matched, but also the spectrum of each component must be matched, particularly the low-frequency end (Armitt and Counihan 1968). Several vertical profiles of mean velocity, turbulence intensity and

Reynolds stress need to be measured to document the boundary layer behavior and to insure similarity to the target atmospheric boundary layer (Snyder 1981). In addition, spectral measurements of turbulence are needed to determine whether the spectra are appropriate for the simulation.

5.5.2 Physical Modelling Facilities in Alberta and their Application

The results of the following studies have been reported elsewhere in this book. The focus here is on the facility itself and the particular type of application.

Yahalom (1972) described a physical modelling study done in the University of Calgary laminar wind tunnel. A 1:250,000 model of the Bow River valley region was exposed to a uniform as well as shear flow of about 0.3 m/s. Dynamic and transport similarity between the model and actual flow was based on analogy between molecular and turbulent transport. In a very slow laminar flow, the molecular viscosity and molecular diffusivity determine the dynamics and the dispersion in the flow whereas in the atmosphere eddy viscosity and turbulent diffusivity are dominant. Reynolds number equivalencies were established based on the relative values of molecular/turbulent viscosity and diffusivity. An artificial shear layer was generated upstream of the study zone by a grid of horizontal copper tubes.

The large wind tunnel at the University of Alberta occupies two floors of the Mechanical Engineering building and has been the site of a variety of studies on flow and dispersion. The tunnel can operate at very low wind speeds which makes it suitable for air pollution studies. Triangular turbulence generators and surface roughness elements are used to simulate the actual atmospheric boundary layer characteristics. The result is a fully developed, atmospheric shear layer up to 1 m thick in about one half of the 11-m test section. Any neutrally stable, adiabatic boundary layers may be simulated. Wilson (1979a) studied the effect of the location and height of the Syncrude tailings pond dike on transport and diffusion of low-level sources within an oil sands plant and of the plant's main stack plume. Peak levels of time-varying concentration and the dike wake effects were investigated.

A smaller wind tunnel facility at the University of Alberta has also been used in a number of studies. This wind tunnel has a square test section with height and width of 30.5 cm. The facility is shown in Figure 5.22. In order to produce a thick shear flow over

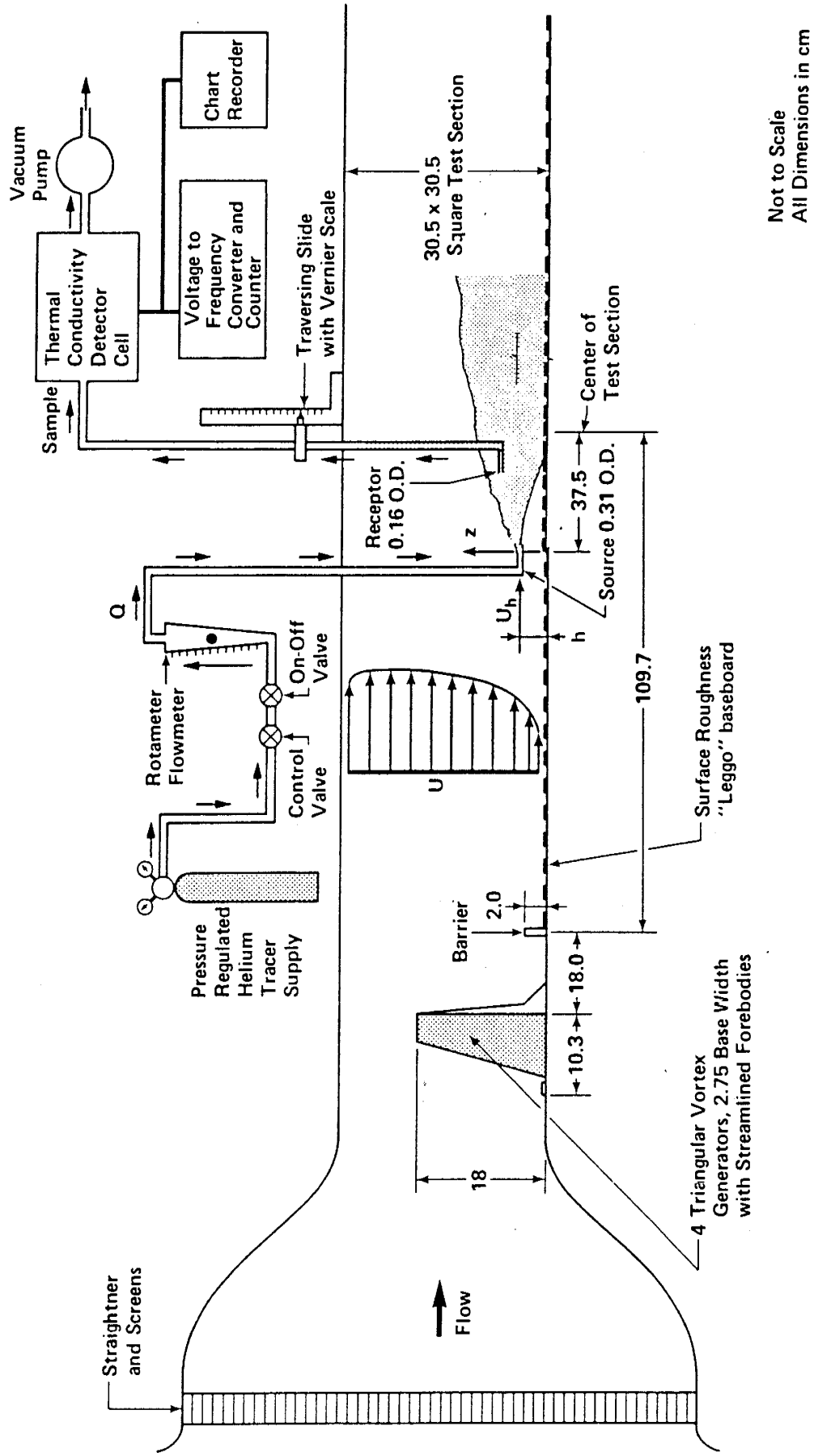


Figure 5.22 University of Alberta small wind tunnel (Netterville 1979).

the surface roughness on the wind tunnel floor, four triangular vortex generators and a barrier are used. Helium tracer gas was used to simulate the plume and concentration measurements can be taken using a movable probe. This facility was used by Wilson (1977a) to investigate plume diffusion over two-dimensional hills and by Netterville (1979) to study the problem of concentration fluctuations in plumes. Hot-wire anemometers were used to determine the turbulence parameters and a wind tunnel reference speed of 10 m/s was chosen, high enough to achieve Reynolds number independence. Helium was used as a tracer, and a specially designed, fast-response, hot-film concentration detector measured the concentration fluctuations.

The dispersion characteristics of a plume as it interacts with a building, have been studied in a simulated boundary layer typical of a lightly built-up urban area. A model building was placed in the test section and a helium tracer gas was measured at various points on and around the building. Reynolds numbers were large enough to assure that flow patterns were in the zone of Reynolds number independence. The results of these simulations for a wide variety of building/plume configurations are given in Wilson and Netterville (1976, 1978) and Wilson and Britter (1982). These studies are reported in Wilson (1976; 1977b, c), Wilson and Chui (1985, 1987), Wilson and Winkel (1982) and Chui and Wilson (1988).

Another physical modelling facility at the University of Alberta, a water channel, was used to study the cross-wind profiles of conditionally averaged mean concentration in plumes (concentration fluctuation problem). The plume from a ground-level source was simulated by the isokinetic release of a neutrally buoyant salt-water-ethanol mixture from a tube on the bed of the water channel. A fully developed, rough wall, open channel flow was used to simulate a neutrally stable atmospheric boundary layer, with surface roughness formed by "Lego" base plate, preceded by a trip fence. The mean and turbulence velocity profiles were measured using a laser doppler anemometer; a fast-response electrical conductivity probe was used to measure concentration. The results are reported in Wilson and Bara (1985) and Bara (1985).

Another water channel at the University of Alberta was used to determine flow patterns and plume/building wake interactions. The channel had a test section width of 67 cm, and was operated at a depth of 30 cm for low-rise buildings and 43 cm for taller buildings. The variation of wind speed and turbulence with height in the atmosphere was

simulated by locating a 1-cm high wall and a row of triangular spikes and surface roughness upstream of the model. Building models had sharp leading edges to encourage Reynolds number insensitivity. Flow patterns were made visible by injecting colored dye through a movable probe which was moved to various positions in the building. The primary purpose of this study was to provide some simple design guides to engineers regarding the placement and design of vents/stacks which are influenced by building effects. The work is reported in Wilson (1979b).

5.6 SUMMARY

Channelling of air flow occurs in the Athabasca River valley north of Fort McMurray as a result of large-scale terrain features, and in southwestern Alberta as a result of a mountain pass. Flow separation occurs in the Bow River valley west of Calgary, creating a recirculation zone. An isolated hill near Pincher Creek causes the wind to speed up by as much as 60% at the crest. Plumes flowing over such hills are squashed by the mean flow, but enhanced turbulence leads to increased ground-level concentrations. Gravity waves occur in the lee of the Rockies more than one-third of the year. Anabatic and katabatic winds have been observed in five different valleys across Alberta. Nocturnal drainage winds in these valleys have maximum wind speed up to 7 m/s located near the mid-point of the valley cross section. Turbulence levels are larger than over flat terrain under inversion conditions. Ground-level concentrations vary with the development of the locally driven flow.

Five prognostic and three diagnostic wind field models have been applied to various complex terrain regions of Alberta. Wind speeds can often be simulated within 10% and wind directions within 20°. Gaussian plume diffusion models can incorporate terrain in several ways, the three most practical being: potential flow solutions, plume spread enhancements and plume height adjustments. Plume spread observations over complex terrain in Alberta show that the vertical and horizontal sigmas are affected differently with enhancement factors averaging about 1.2 for the vertical and 1.7 for the horizontal. In stable conditions, a large terrain obstacle creates two separate flow regimes: above the dividing streamline, a plume will move over the obstacle following potential flow streamlines; below the dividing streamline, a plume will tend to move around the side of the obstacle. Five numerical models of complex terrain diffusion have been applied to various Alberta situations. Predicted concentrations are typically within about 25% of observed concentrations.

As a rule of thumb, when a plume impinges on a building, the concentration on all building surfaces is the average between the rooftop and ground-level concentrations calculated in the absence of the building. Generally, a plume will escape the influence of the building completely if its height is more than 2.5 times the building height. The plume will escape the cavity if its height is more than 1.5 times the building height. The building influence can be incorporated into the Gaussian plume model by means of plume spread correction or by separating the emissions into two plumes, one originating in the cavity. Physical modelling can be used effectively because the turbulence associated with building clusters is near-neutral.

Physical modelling has been used in Alberta to study the flow over mountains, ridges and hills, the flow around industrial complexes and buildings, and the details of concentration fluctuations. Although some limitations are imposed by scaling requirements, physical modelling can be applied to complex situations which would be very difficult to model numerically. Alberta has facilities for this type of work at both the University of Alberta and the University of Calgary.

5.7 REFERENCES

- Alberta Environment. 1988. User's manual for Alberta Environment dispersion models. Standards and Approvals Division. 652 pp.
- Alberta Environment. 1980. Guidelines for plume dispersion calculations. Standards and Approvals Division, Alberta Environment, Edmonton, Alberta. 112 pp.
- Allwine, K.J. and C.D. Whiteman. Green River Air Quality Model Development. 1985. MELSAR: a mesoscale air quality model for complex terrain. Prepared for the U. S. Environmental Protection Agency under a Related Services Agreement with the U. S. Dept. of Energy. Contract DE-AC06-76RLO 1830. Vol. 1. Overview, technical description and users guide EPA 60018-85/017a, 156 pp. Vol. 2. Appendices EPA 60018-85/017b, 359 pp.
- Anderson, G.E. 1971. Meso-scale influences on wind fields. *J. Appl. Meteor.* 10: 377-386.
- Angle, R.P. and P. Gourlay. 1983. Dispersion climatology of the Moose Mountain area. Alberta Environment, February 1983. 58 pp.
- Angle, R.P., C.S. Liu and K. Godard. 1987. Regulating oxides of nitrogen from compressor stations. CPANS/APCA Technical Meeting: New directions in air quality standards, 10 Feb. 1987, Calgary. 23 pp.
- Armitt, J. and J. Counihan. 1968: The simulation of the atmospheric boundary layer in a wind tunnel. *Atmos. Env.* 2 pp. 49-71.
- Arya, S.P.S. and Shipman, M.S. 1979. A model study of boundary layer flow and diffusion over a ridge. 4th Symp. on Turbulence, Diffusion and Air Pollution, Reno, Nevada, 15-18 Jan. pp. 584 - 591. Amer. Meteor. Soc. Boston, Mass.
- Atkinson, B.W. 1981. Meso-scale atmospheric circulations. Academic Press, Toronto. 495 pp.
- Avissar, R., M.D. Moran, G. Wu, R.N. Meroney, and R.A. Pielke. 1990. Operating ranges of mesoscale numerical models and meteorological wind tunnels for the simulation of sea and land breezes. *Boundary Layer Meteorology*, 50:227 - 275.
- Ayer, H.S. 1961. On the dissipation of drainage wind system in valleys in morning hours. *J. Meteor.* 18(4): 560 - 563.
- Bara, B. 1985. Water channel simulation of concentration fluctuations from a ground level source. M.Sc. Thesis. Dept. of Mech. Eng., University of Alberta. 170 pp.
- Barr, S. and M.M. Orgill. 1989. Influence of external meteorology on nocturnal valley drainage winds. *J. Appl. Meteor.* 28: 497 - 517.
- Benjamin, S.F., J.J. Havlena, R.D. Rowe. 1978. Field investigations of the dispersion of pollution across a ridge. Proceedings of the 9th International Technical Meeting on Air Pollution Modelling and Its Application, Toronto, Ontario, Canada. 45-54.

- Benjamin, S.F., J.J. Havlena, R.D. Rowe. 1977. Observational studies of air quality and air flow over a ridge downwind of a stack. Proceedings of the 70th Annual Meeting of the Air Pollution Control Association. Toronto, Canada, June 20 - 24, 1977. paper No. 77-4.3. 16 pp.
- Bergen, J.D. 1969. Cold air drainage on a forested mountain slope. *J. Appl. Meteor.* 8: 884 - 895.
- Bradley, E.F. 1980. An experimental study of the profiles of wind speed, shearing stress and turbulence at the crest of a large hill. *Quart. J. Roy. Meteor. Soc.*, 106, 101 - 123.
- Briggs, G.A. 1975. Diffusion estimates for small emissions in Atmospheric Turbulence and Diffusion Laboratory 1973 Annual Report, Publication No. ATDL-106, pp. 83-145. NOAA, Oak Ridge, Tenn.
- Buettner, K.J.K. and N. Thyer. 1966. Valley winds in the Mount Ranier area. *Arch. Meteor. Geophys. Bioklim., Ser. B*, 14: 125 - 147.
- Cermak, J.E. 1975, Applications of fluid mechanics to wind engineering. Presented at Winter Annual Meeting of the ASME, New York. Nov. 17-21, Freeman Scholar Lecture, American Society of Mechanical Engineers, New York, 1974. *Trans. ASME J. Fluids Eng.* 97:9-38.
- Cermak, J.E., D.J. Lombardi and R.S. Thompson. 1974. Applications of physical modelling to investigations of air pollution problems in urban areas. 67th Annual Meeting of the Air Pollution Control Assoc., June 9-13, Paper No. 74-160. Denver, Colorado.
- Chui, E.H. and D.J. Wilson. 1988. Effect of Varying Wind Direction on Exhaust Gas Dilution, *Journal of Wind Engineering and Industrial Aerodynamics* 31, pp. 87-104.
- Chung, K.P. 1979. Tetroon flights across a two-dimensional ridge under stable atmospheric conditions. M. Sc. Thesis, University of Calgary, Alberta. 166 pp.
- Chung, K.P. and R.D. Rowe. 1979. Tetroon flights across a ridge under stack atmospheric conditions. Dept. of Chemical Engineering, University of Calgary, Calgary, Alberta. 118 pp.
- Clements, W.E., J.A. Archuleta and D.E. Hoard. 1989. Mean structure of the nocturnal drainage flow in a deep valley. *J. Appl. Meteor.* 28: 457 - 462.
- Committee Appointed by the Electricity Commissioners. 1932. The measures which have been taken in this country and in others to obviate the emission of soot, ash, grite and gritty particles from the chimneys of electric power stations. Her Majesty's Stationary Office, London, England.
- Coulter, R.L., M. Orgill and W. Porch. 1989. Tributary fluxes into Brush Creek Valley. *J. Appl. Meteor.* 28: 555-568.
- Danard, M.B. and J. Galbraith. 1987. Meso gamma wind modelling for the 1988 Calgary Winter Olympics. Part I - Model description and data used. Prep. by Atmospheric dynamics Corporation for the Atmos. Environ. Ser. Co. No. KM643-6-0447/01-SG, File No. 03SG. 23294-6-0447.

- Danard, M.B. and M. Gray. 1982. Modelling topographic effects on winds in the Alberta Oil Sands area. AOSERP Rep. No. 126, prepared by Atmospheric Dynamics Corporation for Research Management Division, Alberta Environment, Edmonton, Mar. 1982, 89 pp.
- Davison, D.S., R. Rudolph and R.B. Lantz. 1985. A particle dispersion model for real-time assessment and contingency planning of chemical accidents. In proc. of 1985 joint annual meeting of CPANS and PNWIS of the Air Poll. Cont. Assoc., Calgary, Alberta. Nov. 13-15. paper F5.
- Davidson, B. and P.K. Rao. 1963. Experimental studies of the valley- plain wind. *Int. J. Air Wat. Poll.* 7: 907 - 923.
- Defant, F. 1951. Local winds: In *Compendium of Meteorology*. American Meteorological Society, Boston, Massachusetts. pp. 655 - 672.
- Defant, F. 1949. Zur theorie der hangwinde, nebst bemerkungen zur theorie der berg-und talwinde. (A theory of slope winds, along with remarks on the theory of mountain winds and valley winds). *Arch. Meteor. Geophys. Bioklim. Ser. A*, 1, 421 - 450. [English translation: Whiteman, C.D. and E. Dreiseitl, 1984. *Alpine Meteorology. Transitions of classic contributions by A. Wagner, E. Ekhardt and F. Defant. PNL-5141/ASCOT-84-3. Pacific Northwest Laboratory, Richland, Washington, 121 pp*].
- Dickerson, M.H. Editor. 1980. Atmospheric studies in complex terrain (ASCOT) - information survey. Lawrence Livermore National Laboratory, University of California, Livermore, California 94550. UCID-18572, ASCOT-80-3. 247 pp.
- Dicristofaro, D.C., D.G. Strimaitis, B.R. Greene, R.J. Yamartino, and A. Venkatram. 1985. EPA complex terrain model development, Fifth milestone report - 1985. EPA/600/3-85/069. 309 pp.
- Doran, J.C. and T.W. Horst. 1981. Velocity and temperature oscillations in drainage winds. *J. Appl. Meteor.* 20: 361-364.
- Drazin, P.G. 1961. On the steady flow of a fluid of variable density past an obstacle. *Tellus* 13: 239 - 251.
- Egan, B.A. 1975. Turbulent diffusion in complex terrain. Chapter 4 of *Lectures on Air Pollution and Environmental Impact Analyses*. D. Haugen (ed). American Meteorological Society. Boston, MA. 112-135.
- Electric Power Research Institute (EPRI). 1982. Survey of plume models for atmospheric application. Prepared by Systems Applications Inc, EPRI EA-2243, Project 1616-9. 133 pp.
- Energy Resources Conservation Board (ERCB) and Alberta Environment. 1988. Informational Letter IL 88-5. Application for approval of Natural-Gas-Driven Compressors. 9 May 1988. 8 pp.
- Engineering-Science (1980) "Field Validation of Atmospheric dispersion models for natural gas compressor stations". Prepared for Pipeline Research Committee American Gas Association, PR-133-78.

- Engineering-Science (1977) Workbook for predicting atmospheric dispersion of engine exhaust emissions from natural gas compressor stations (Preliminary Version). Am. Gas Assocn., Project PR-133-78. 73 pp.
- Environment Ontario. 1987. Air Pollution Regulation 308, Appendix H - MOE Application. 170 pp.
- Fabrick, A.J., R.C. Sklarew and J.C. Wilson. 1977. Point source modelling: state of the art assessment of computer simulation of localized air pollutant emissions. Form and Substance Inc. Westlake Village, California. 197 pp.
- Freytag, C. 1987. Results from the MERKUR experiment: mass budget and vertical motions in a large valley during mountain and valley wind. Meteor. Atmos. Phys, 37:129 -140.
- Galbraith, J. and M.B. Danard. 1987. Compact Alberta diffusion model user's guide. Report prep for Pollution Control Division of Alberta Dept. of Environment by Atmospheric Dynamics Corp.
- Gandemer, J. 1976. Inconfort du au vent aux abords des batiments: concepts ae'rodynamiques. Cahier du Centre Scientifique et Technique du Batiment, No. 170, Cahier 1384, Centre Scientifique et Technique du Batiment, Nantes, France.
- Gourlay, P., R.P. Angle and S.K. Sakiyama. 1984. Diurnal variation of valley flows in the Alberta foothills. Paper presented at the 18th Annual Congress of the Canadian Meteorological and Oceanographic Society, 29 May-1 June, 1984, Halifax, Nova Scotia. (Abstract).
- Gudiksen, P.H., M.H. Dickerson, R. Lange and J.B. Knox. 1985. Atmospheric field experiments for evaluating pollutant transport and dispersion in complex terrain. Air Pollution Modelling and Its Application IV, C. De Wispelaere (Editor), Plenum Press, New York, 507 - 528.
- Gryning, S.E., L. Mahrt and S. Larsen. 1985. Oscillating nocturnal slope flow in a coastal valley. Tellus 37A: 196-203.
- Hage, K.D. 1981. Wave like oscillations in urban valley drainage flow. Proceedings from the Fifth Symposium on Turbulence, Diffusion and Air Pollution. Mar. 9 - 13, 1981, Atlanta, GA. American Meteorological Society, Boston, MA. 179-180.
- Hage, K.D. 1980. Urban valley air pollution. Proceedings from the Second Joint Conference on Applications of Air Pollution Meteorology, 24 - 28 March, 1980, New Orleans, La. American Meteorological Society, Boston, MA. 515-519.
- Hage, K.D. 1979. Air pollution study of the North Saskatchewan River valley in Edmonton, Alberta. Final Report, Research Proposal RS77-6, Research Secretariat, Alberta Environment, Edmonton. 90 pp.
- Halitsky, J. 1968. Gas diffusion near buildings, in Meteorology and Atomic Energy - 1968, D.H. Slade (ed), Chap. 5-5 pp. 221-255, USAEC Report TID-24190, Environmental Science Services Administration, NTIS.

- Halitsky, J. 1962. Diffusion of vented gas around buildings. *J. Air Poll. Cont. Assoc.* 12(2): 74 - 80.
- Halitsky, J., G.A. Magony and P. Halpern. 1965. Turbulence due to topographical effects. Geophysical Sciences Laboratory Report, TR-65-2. Depart. of Met. and Ocean, New York University. 75 pp.
- Hawkins, J.E. and G. Nonhebel. 1955. Chimneys and the dispersal of smoke. *Jour. Inst. Fuel* 28, 530-545.
- Hewson, E.W. and G.C. Gill. 1944. Meteorological investigations in Columbia River Valley near Trail, British Columbia - Part II. Report submitted to the Trail Smelter Arbitral Tribunal, Bulletin 453, U. S. Bureau of Mines, Washington, D.C. 228 pp.
- Hoffnagle, G.F., B.A. Egan and B. R. Greene. 1977. Application of models in complex terrain. Joint Conf. on Appl. of Air Poll. Meteorology. Salt Lake City, Utah. American Meteorological Society, Boston, MA. 196-203.
- Hosker, R.P. 1984. Flow and diffusion near obstacles. Chapter 7 of *Atmospheric Science and Power Production*. D. Randerson Editor. USDOE, DOE/TIC-27601. (DE 84005177). 241-326.
- Hosker, R.P. Jr. 1982. A survey of the utility of laboratory modelling of atmospheric boundary layer dispersion and related phenomena. In workshop on the parameterization of mixed layer diffusion, October 1981. NOAA, ATDL contribution file No. 81/19.
- Hoydysh, W.G., R.A. Griffiths and Y. Ogawa. 1974. A scale model study of the dispersion of pollution in street canyons. 67th Annual Meeting of the Air Pollution Control Assoc., June 9-13, Paper No. 74-157. Denver, Colorado.
- Huber, A.H. 1984. Evaluation of a method for estimating pollution concentrations downwind of influencing buildings. *Atmos. Environ.* 18 (11): 2313 - 2338.
- Huber, A.H. 1977 Incorporating building/terrain wake effects on stack effluents. Joint Conference on Applications of Air Pollution Meteorology. Salt Lake City, Utah. American Meteorological Society, Boston, MA. 353-356.
- Hunt, J.C.R. and W.H. Snyder. 1980. Experiments on stably and neutrally stratified flow over a model three-dimensional hill. *J. Fluid Mech.* 96: 671 - 704.
- Hunt, J.C.R., J. S. Puttock and W.H. Snyder. 1979. Turbulent diffusion from a point source in stratified and neutral flows around a three-dimensional hill; Part I: diffusion equation analysis. *Atmos. Environ.* 13: 1227 - 1239.
- Hunt, J.C.R., W.H. Snyder and R.E. Lawson, Jr. 1978. Flow structure and turbulent diffusion around a three-dimensional hill; fluid modelling study on effects of stratification; Part I - flow structure. EPA-600/4-78-041, U. S. Environmental Protection Agency, Research Triangle Park, NC. 96 pp.

- Hunt, J.C.R. and P.J. Mulhearn. 1973. Turbulent dispersion from sources near two-dimensional obstacles. *J. Fluid Mech.* 61: 245 - 274.
- Intercomp, no date. The effect on air quality of the Athabasca synthetic crude plants, Intercomp Report. 13 pp.
- Intercomp. 1972. Air quality study, Petrogas plant, Balzac, Alberta. Intercomp Report to Petrogas Processing.
- Kao, T.W. 1965. The phenomena of blocking in stratified flow. *J. Geophys. Res.* 70: 815 - 822.
- Kessler, R.C., M.A. Yocke and R.E. Morris. 1987. Designing a wind field module. Systems Applications Inc., San Rafael, CA.
- Klassen, W. 1962. Micrometeorological studies in the North Saskatchewan River valley at Edmonton. Meteorological Branch, Canada, Dept. of Transport, CIR-3652, TEC-408. 16 pp.
- Lange, R. 1978. ADPIC - A three-dimensional particle-in-cell model for the dispersal of atmospheric pollutants and its comparison to regional tracer studies. *J. of Appl. Meteor.* Vol. 17, pp. 320-329.
- Lantz, R.B. 1972. Application of a three-dimensional numerical model to air pollutant calculations. Paper 72-141, 65 Annual Meeting of APCA, 1972. June 18-22, Miami Beach.
- Lavery, T.F., D.G. Strimaitis, A. Venkatram, B.R. Greene, and D.C. DeCristofaro. 1983. EPA Complex terrain model development. Third milestone report - 1983. EPA-600/3-83-101. 293 pp.
- Leahey, D.M. and M.C. Hansen. 1982. Influences of terrain on plume level winds in the Athabasca Oil Sands area. *Atmos. Environ.* 16(12): 2849 - 2854.
- Leahey, D.M., D.J. Wilson, S.W. Behie, J.B. Railton. 1979 Evaluation of downdraft effects from a large thermal plant building. PNWIS-APCA meeting, Edmonton, Alberta. paper 79-4.1.
- Leahey, D.M. and M.J.E. Davies. 1978. Bivane studies of air flow over irregular terrain. Proceedings of the 9th NATO International Technical Meeting on Air Pollution and its Application. Toronto. pp. 89 - 100.
- Leahey, D.M. 1974. A study of air flow over irregular terrain. *Atmos. Environ.* 8:783 - 791.
- Leahey D.M. and R.D. Rowe. 1974. Observational studies of atmospheric diffusion processes over irregular terrain. Paper 74 - 67 presented at the 67th meeting of the Air Poll. Cont. Assoc. June 9-13. Denver.
- Leahey, D.M. and J. Halitsky. 1973. Low wind turbulence statistics and related diffusion estimates from a site located in the Hudson River Valley. *Atmos. Environ.* 7: 49 - 61.

- Leahey, D.M. and H.S. Hicklin. 1973. Tetroon studies of diffusion potential in the airshed surrounding the Crowsnest Pass area. *Atmosphere* 11(3): 77 - 87.
- Lee, C.Z. 1979. Plume trajectories across a ridge under stable atmospheric conditions. M. Sc., Thesis, University of Calgary, Alberta. 145 pp.
- Lenschow, D.H., B.B. Stankov and L. Mahrt. 1979. The rapid morning boundary-layer transition. *J. of Atmos. Sci.* 36: 2108 - 2124.
- Lester, P.F. 1976a. Evidence of long lee waves in southern Alberta. *Atmosphere*, 14: 28 - 36.
- Lester, P.F. 1976b. A lee wave cloud climatology for Pincher Creek, Alberta. Environmental Sciences Centre (Kananaskis). University of Calgary, Calgary, Alberta, T2N 1N4. 27 pp.
- Lester, P.F. and J.I. MacPherson. 1976. Waves and turbulence in the vicinity of a chinook arch cloud. Environmental Sciences Centre (Kananaskis). University of Calgary, Calgary, Alberta, T2N 1N4, 30 pp.
- Longley, R.W. 1969. The diurnal pressure wave in western Canada. *J. Appl. Meteor.* 8: 754 - 761.
- Longley, R.W. 1968. The diurnal variation of wind direction at Calgary. *Atmosphere*. 6: 23 - 29.
- Lott, R.A. 1986. Model performance - plume dispersion over elevated terrain. *Atmos. Environ.* 20(8): 1547 - 1554.
- Ludwig, F.L. 1984. Transport and diffusion calculations during time- and space-varying meteorological conditions using a microcomputer. *J. Air Poll. Cont. Assoc.* 34: 227 - 232.
- Ludwig, F.L. and G. Byrd. 1980. An efficient method for deriving mass-consistent flow fields from wind observations in rough terrain. *Atmo. Environ.* 14: 585-587.
- MacHattie, L.B. 1968. Kananaskis valley winds in summer. *J. Appl. Meteor.* 7(1): 348 - 352.
- Manins, P.C. and B.L. Sawford. 1979. Katabatic winds: a field case study. *Quart. J. Roy. Meteor. Soc.* 105: 1011 - 1025.
- Mason, P.J. and R.I. Sykes. 1979. Flow over an isolated hill of moderate slope. *Q.J.R. Meteor. Soc.* 105: 383 - 395.
- McVehil, G.E. 1988. Review of U.S EPA plume dispersion models. Report prep. for Western Research, Calgary, Alberta. 55 pp.
- MEP/Envirocon. 1976. August 1975 and February 1976 plume dispersion studies at the Coleman gas processing plant to comply with Licence No. 75-AL-021. Prepared for the Saratoga Gas Processing Co. Ltd., Coleman, Alberta. 82 pp.

- Meroney, R.N. and D.E. Neff 1979. Laboratory simulation of liquid natural gas vapor dispersion over land and water. In Proc. of 5th Internat. Conf. on Wind Engineering. Ft. Collins Co. Vol. 2. J.E. Cermak, ed. Pergamon Press, New York, 1980. 1139-1150.
- Meroney, R.N., A.J. Bowen, D. Lindley and J. Pearse. 1978. Wind characteristics and field measurements in Rakaia Gorge, New Zealand. Dept. of Energy Report RLO/2438-77/2, 219 pp.
- Mickle, R.E., J.R. Salmon and P.A. Taylor. 1984. Kettles Hill, 1984: velocity profile measurements over a low hill. Report, AQRB-84-012-L, Atmospheric Environment Service. Downs-view (Toronto), Canada. October 1984. 66 pp.
- Moninger, W.R., W.L. Eberhard, G.A. Briggs, R.A. Kropfli and J.C. Kaimal. 1983. Simultaneous radar and Lidar observations of plumes from continuous point sources. 21st Radar Meteorology Conference, Amer. Meteor. Soc., Boston. MA. 246-250.
- Muller, H. and C.D. Whiteman. 1988. Breakup of a nocturnal temperature inversion in the Dischma Valley during DISKUS. J. Appl. Meteor. 27: pp. 188 - 194.
- Munn, R.E. 1966. Descriptive Micrometeorology. Academic Press, New York, 245 pp.
- Neff, W.D. and C.W. King. 1987. Observations of complex-terrain flows using acoustic sounders: experiments, topography and winds. Bound. Layer Meteor. 40: 363-392.
- Netterville, D.D.J. 1979. Concentration fluctuations in plumes. Syncrude Canada Environmental Research Monograph 1979-4. 228 pp.
- Nicholls, J.M. 1973. The airflow over mountains. Research. 1957 - 1972. WMO Technical Note No. 127, WMO No. 355, Geneva. 73 pp.
- Ogawa, Y., R. Griffiths, and W.G. Hoydysh. 1975. A wind-tunnel study of sea breeze effects. Boundary-Layer Meteorol. 8(2), pp 141-161.
- Orgill, M.M., J.E. Cermak and L.O. Grant. 1971. Laboratory simulation and field estimates of atmospheric transport - dispersion over mountainous terrain. Report CER70-71MMO-JEC-LOG40, Fluid Dynamics and Diffusion Laboratory, College of Engineering, Colorado State University, Fort Collins, Colorado. 143 pp.
- Paterson, R.D. and K.D. Hage. 1979. Micrometeorological study of an urban valley. Boundary Layer Meteor. 17: 175 - 186.
- Pielke, R.A. 1984. Mesoscale meteorological modeling. Academic Press, Inc. Orlando, Florida 32887. 612 pp.
- Pielke, R.A., R.T. McNider, M. Segal and Y. Mahrer. 1983. The use of a mesoscale numerical model for evaluations of pollutant transport and diffusion in coastal regions and over irregular terrain. Bull. Amer. Meteor. Soc., 64: 243 - 249.
- Porch, W.M., R.B. Fritz, R.J. Coulter and R.H. Gudiksen. 1989. Tributary, valley and sidewall air flow interactions in a deep valley. J. Appl. Meteor. 28: 578 - 589.

- Raupach, M.R. and A.S. Thom. 1981. Turbulence in and above plant canopies. In *Ann. Rev. Fluid Mech.* 13 (Annual Reviews, Inc.) Palo Alto, CA, pp 97-129.
- Reid, J. D., J. L. Walmsley, I. B. Findleton and A. D. Christie. 1979. An assessment of the models LIRAQ and ADPIC for application to the Alberta Oil Sands area. Alberta Environment, Research Management Division, AOSERP Report 66. 95 pp.
- Rowe, R.D. and I. Tas. 1986. A simple technique for stack design in complex terrain. Fifth Joint Conf. on Applications on Air Poll. Meteorology. Nov. 18 - 21, 1986 Chapel Hill, N. Carolina. American Meteorological Society, Boston, MA. 207-210.
- Rowe, R.D., S.F. Benjamin, K.P. Chung, J.J. Havlena and C.Z. Lee. 1982. Field studies of stable air flow over and around a ridge. *Atmos. Environ.* 16: 643 - 653.
- Rowe, R.D. 1980. A simple two-layer model for stable air flow across terrain features. *Proceedings of the 2nd Joint Conference on Applications of Air Pollution Meteorology*, New Orleans. 559 - 566.
- Rowe, R.D., C.Z. Lee and J.J. Havlena. 1980. The elevation of stable atmosphere plumes above a ridge. *Proceedings of the 73rd Annual APCA Meeting*, June 22-27, Montreal. Paper No. 80-55.5.
- Rudolph, R.C. 1980. A numerical model of pollutant transport in a small urban valley. Department of Geography, University of Alberta, Edmonton. M. Sc. Thesis. 126 pp.
- Rudolph, R. and K.D. Hage. 1983. Potential flow model of pollutant recirculation in a small valley. *Arch. Met. Geoph. Biocl. Ser. B* 33:175-185.
- Sakiyama, S.K. 1990. Drainage flow characteristics and inversion breakup in two Alberta mountain valleys. *J. Appl. Meteor.* 29: 1015 - 1030.
- Sakiyama, S.K. 1983. Numerical modelling and field verification for the Shell Moose Mountain Sour Gas Project. Alberta Environment. 98 pp.
- Sakiyama, S.K. and R.P. Angle. 1988. Modelling tracer study for risk assessment of a proposed sour gas facility. *Env. Monitor and Assess.* 10(2): 133 - 146.
- Sakiyama, S.K., P. Freeman and R.P. Angle. 1988. A dispersion model for use in the design of compressor station stacks. Presented at the CPANS - APCA annual meeting. May, 1988, Calgary, Alberta.
- Sakiyama, S.K., R.P. Angle and P.R. Gourlay. 1984. Vertical structure of mountain valley flows in Alberta. Paper presented at the 18th Annual Congress of the Canadian Meteorological and Oceanographic Society. 29 May to 1 June, 1984, Halifax, Nova Scotia. (Abstract).
- Salmon, J.R., H.W. Teunissen, R.E. Mickle, and P.A. Taylor. 1988. The Kettles Hill project; field observations, wind-tunnel simulations and numerical model predictions for flow over a low hill. *Boundary-Layer Meteorol.* 43(4) pp. 309-343.

- Sawford, B.L. 1982. Lagrangian Monte Carlo simulation of the turbulent motion of a pair of particles. *Quart. J. Roy. Meteor. Soc.* 108: 207 - 213.
- Scire, J.S. and A. Bass. 1983. COMPLEX/PFM air quality model user's guide. EPA-600/S8-83-015, U. S. Environmental Protection Agency, Research Triangle Park, NC. 115 pp.
- Scorer, R. and A. Verkaik. 1989. *Spacious skies*. David and Charles Publishers, London. 192 pp.
- Seinfeld, J.H. 1988. Ozone air quality models: a critical review. *J. Air Poll. Cont. Assoc.* 38(5): 616 - 645.
- Sheppard, P.A. 1956. Airflow over mountains. *Quart. J. Roy Meteor. Soc.* 82: 528 - 529.
- Sherlock, R.H. and E.J. Leshner. 1954. Role of chimney design in dispersion of waste gases. *Air Repair* 4, (2):13-23.
- Slawson, P.R., G.A. Davidson and C.S. Maddukurri. 1979. Dispersion modelling of a plume in the tar sands area. *Synchrude Env. Research Report* 1980-1, 316 pp.
- Slawson, P.R., G. Davidson, W. McCormick and G. Raithby. 1978. A study of dispersion characteristics of the G.C.O.S. plume. prepared by Envirodyne Ltd. for Synchrude Canada Ltd. 113 pp.
- Smith, M.E. (Editor). 1968. Recommended guide for the prediction of the dispersion of airborne effluents. American Society of Mechanical Engineers, New York, N.Y. 85 pp.
- Snyder, W.H., R.S. Thompson, R.E. Eskridge, R.E. Lawson, I.P. Castro, J.T. Lee, J.C.R. Hunt and Y. Ogawa. 1985. The structure of a strongly stratified flow over hills: The dividing streamline concept. *J. Fluid Mech.*, 152, 249-288.
- Snyder, W.H. and J.C.R. Hunt. 1984. Turbulent diffusion from a point source in stratified and neutral flows around a three dimensional hill - II Laboratory measurements of surface concentrations. *Atmos. Environ.* 18(10): 1969 - 2002.
- Snyder W.H. 1981. Guideline for fluid modelling of atmospheric diffusion (Fluid modelling rept. No. 10). EPA-600/8-81-009. Research Triangle Park, NC. 200 pp.
- Snyder, W.H. and R.E. Lawson, Jr. 1976. Determination of a necessary height for a stack close to a building - a wind tunnel study. *Atmos. Environ.* 10: 683-691.
- Snyder W.H. 1972. Similarity criteria for the application of fluid models to the study of air pollution meteorology. *Bound. Layer Meteor.* 3: 113 - 134.
- Start, G.E., C.R. Dickson and L.L. Wendell. 1975. Diffusion in a canyon within rough mountainous terrain. *J. Appl. Meteor.* 14: 333 - 346.
- Start, G.E., C.R. Dickson, L.L. Wendell. 1973. Diffusion in a canyon within rough mountainous terrain. NOAA Tech. Memo. ERL - ARL 38, Air Resources Laboratories, Idaho Falls, Idaho. 43 pp.

- Stone, G.L. and D.E. Hoard. 1989. Low-frequency velocity and temperature fluctuations in katabatic valley flows. *J. Appl. Meteor.* 28: 477 - 488.
- Strimaitis, D.G., R.J. Paine, B.A. Egan, R.J. Yamartino and B. Greene. 1988. EPA complex terrain model development. Final Report. EPA 1600/3-88/006, USEPA, Research Triangle Park, NC. 485 pp.
- Strimaitis, D.G., T.F. Lavery, A. Venkatram, D.C. Dicristofaro, B.R. Greene, and B.A. Egan. 1985. EPA complex terrain model development. Fourth milestone report - 1984. EPA Report No. EPA -600/3-84/110. 345 pp.
- Strimaitis, D.G., J.S. Scire and A. Bass. 1983. COMPLEX/PFM air quality model User's guide. EPA-600/8-83-015, U. S. Environmental Protection Agency, Research Triangle Park, NC. 117 pp.
- Strimaitis, D.G., A. Venkatram, B.R. Greene, S. Hanna, and S. Heisler. 1983. EPA complex terrain model development, Second milestone report - 1982. EPA-600/3-83-015. 402 pp.
- Strom, G.H. 1976. Transport and diffusion of of stack effluents. *Air Pollution*, Vol. 1, 3rd ed. A.C. Stern (ed) Academic Press. 401 - 501.
- Strom, G.H., M.Hackman, and E.J. Kaplin. 1957. Atmospheric dispersal of industrial stack gases determined by concentration measurements in scale model wind tunnel experiments. *J. Air Poll. Control Assoc.* 7: 198 - 203.
- Strom, G.H. 1955-1956. Wind tunnel scale model studies of air pollution from industrial plants. *Ind. Wastes*. Sept. 10, 1955, Nov./Dec. 1955, Jan/Feb. 1956.
- Stumke, H. 1964a. Consideration of simplified terrain types in the calculation of the turbulent propagation of chimney gases. *Staub*. 24: 175 - 182. Translated for Oak Ridge National Laboratory. 21 pp.
- Stumke, H. 1964b. Correction of the chimney height due to an influence of the terrain. *Staub* 24: 525 - 528. Translated for Oak Ridge National Laboratory. 10 pp.
- Taylor, P.A. 1977. Some numerical studies of surface boundary-layer flow above gentle topography. *Bound. Layer Meteor.* 11: 439 - 465.
- Taylor, P.A., R.E. Mickle, J.R. Salmon and H.W. Teunissen. 1983a. The Kettles Hill Experiment - site description and mean flow results. Report AQRB-83-002-L, Atmospheric Environment Service, Downs-view (Toronto), Canada, March 1983.
- Taylor, P.A., J.L. Walmsley and J.R. Salmon. 1983b. A simple model of neutrally stratified boundary layer flow over real terrain incorporating wavenumber-dependent scaling. *Boundary Layer Meteor.* 26: 169 - 189.
- Tesche, T.W., J.L. Haney and R.E. Morris. 1987. Performance evaluation of four grid-based dispersion models in complex terrain. *Atmos. Environ.* 21(1): 233 - 256.

- The Landplan Group. 1984. Mt. Allan, ski area master plan. prep. for the Province of Alberta, Alberta Tourism and Small Business. 162 pp.
- Thomson, R.B., R.P. Angle and S.K. Sakiyama. 1987. Selecting air quality - acid deposition models for mesoscale application. *J. Air Poll. Cont. Assoc.* 37(3): 260 - 265.
- Thorp, J.M. and M.M. Orgill. 1986. A "waterfall" of air. *Weatherwise*. 39:319-322.
- Thyer, N. 1981. Diurnal variations of upper winds in summer in Alberta. *Atmosphere-Ocean* 19(4): 337 - 344.
- Townsend, A.A. 1956. The structure of turbulent shear flow. Cambridge University Press, 315 pp.
- Venkatram, A. 1986. Dispersion in complex terrain. Short course on air pollution modelling. March 17 - 21, 1986. San Diego, CA. American Meteorological Society, Boston, MA.
- Vet, R.J., N.W. Reid, E. Alp. and S.J. Diehl. 1982. Compressor exhaust chemistry and wake diffusion. Report for the Alberta Petroleum Industry Government Environmental Committee. Prepared by Concord Scientific Corporation. Report No. 67-1. 205 pp.
- Vet, R.J., N.W. Reid and G.D. Stuble. 1981. Compressor exhaust chemistry and wake diffusion phase I: An annotated bibliography of chemistry, flow and dispersion in the wake of buildings. Report prepared for the Alberta Petroleum Industry - Government Environmental Committee by Concord Scientific Corporation. Report CSC-67-1. 288 pp.
- Wallis, J., K. Aziz and J.K. Donnelly. 1975. Users' guide for irregular terrain steady-state air pollution program. Chemical Engineering Department, University of Calgary, Calgary, Alberta. 69 pp.
- Walmsley, J.L., P.A. Taylor, and T. Keith. 1986. A simple model of neutrally stratified boundary-layer flow over complex terrain with surface roughness modulations (MS3DJH/3R). *Bound. Layer Meteor.* 36: 157 - 186.
- Walmsley, J.L., P.A. Taylor and R. Mok. 1980. MS3DJH - a computer model for the study of neutrally stratified boundary layer flow over isolated hills of moderate slope. Air Quality and Inter-Environmental Research Branch, Atmospheric Environment Service, 4905 Dufferin St., Downs-view, Ontario. M3H 5T4, Apr. 1980. 43 pp.
- Webb, R. 1982. Final report - Moose Mountain air and water quality monitoring program - prep. for Shell Canada Resources Ltd., 65 pp.
- Wedding, J.B., D.J. Lombardi and J.E. Cermak. 1977. A wind tunnel study of gaseous pollutants in city street canyons. *J. Air. Poll. Control Assoc.*, 27(6): 557-566.
- Western Mesoscale Modelling Task Group. 1985. A review of mesoscale modelling for application to Western Canada. Prepared for the Western Canada LRTAP Technical Committee. 31 pp.

- Western Research and Development. 1978. Assessment of dispersion effects which will influence the Keepphills stack plume - with emphasis on building downdrafts. Ref. 2715-R1. 20 pp. plus Appendix.
- Whiteman, C.D. 1982. Breakup of temperature inversions in deep mountain valleys: part I. observations. *J. Appl. Meteor.* 21: 270 - 289.
- Whiteman, C.D. and S. Barr. 1986. Atmospheric mass transport by along-valley wind systems in a deep Colorado valley. *J. Climate Appl. Meteor.* 25: 1205 - 1212.
- Whiteman, C.D. and T.B. McKee. 1978. Air pollution implications of inversion descent in mountain valleys. *Atmos. Environ.* 12, 2151 - 2158.
- Wilson, D.J. and E.H. Chui. 1987. Effect of Turbulence from Upwind Buildings on Dilution of exhaust gases, *ASHRAE Transactions*, 93 part 2:2186-2197.
- Wilson, D.J. and B. Bara. 1985. Crosswind profiles of conditionally averaged mean concentration in plumes. November 1985, Boulder, Colorado. American Meteorological Society. Extended Abstracts. 7th Symp. on Turb. and Diffusion, paper No. P3.11. 259-261.
- Wilson, D.J. and E.H. Chui. 1985. Influence of Exhaust Velocity and Wind Incidence Angle on Dilution from Roof Vents, *ASHRAE Transactions* 91 part 2B. 1693-1706.
- Wilson, D.J. and R.E. Britter. 1982. Estimates of building surface concentrations from nearby point sources. *Atmos. Env.* 16 (11) pp 2631-2646.
- Wilson, D.J. and G. Winkel. 1982. The Effect of Varying Exhaust Stack Height on Contaminant Concentration at Roof Level, *ASHRAE Transactions* 88, Part 1, pp. 513-533.
- Wilson, J.D., G.W. Thurtell and G.E. Kidd. 1981. Numerical simulation of particle trajectories in inhomogeneous turbulence. I: Systems with constant Turbulent Velocity Scale. *Boundary Layer Meteor.* 21: 295 - 313.
- Wilson, D.J. 1980. Influence of a hill ridge on ground level concentration of stack gas plumes. *Env. Tech. Letters* 1: 177 - 184.
- Wilson, D.J. 1979a. Wind tunnel simulation of plume dispersion at Syncrude Mildred Lake site. *Syncrude Canada Environmental Research Monograph* 1971-1, 198 pp.
- Wilson, D.J. 1979b. Flow patterns over flat-roofed buildings and application to exhaust stack design *A.S.H.R.A.E. Trans.* 85 Part 2, pp. 284-295.
- Wilson, D.J. and D.D.J. Netterville. 1978. Interaction of a roof level plume with a downwind building. *Atmos. Env.* 12, 1051-1059.
- Wilson, D.J. 1977a. Plume diffusion over downwind two dimensional hills: a wind tunnel study. Dept. of Mechanical Engineering, University of Alberta, Edmonton, Alberta. Report to Alberta Environment, Research Project #75-8, 105 pp.

- Wilson, D.J. 1977b. Effect of Vent Stack Height and Exit Velocity on Exhaust Gas Dilution, Trans. ASHRAE 83, pp. 157-166.
- Wilson, D.J. 1977c. Dilution of Exhaust Gases From Building Surface Vents, Trans. ASHRAE 83, pp. 168-176.
- Wilson, D.J. 1976. Contamination of Air Intakes from Roof Exhaust Vents. ASHRAE 82, pp. 1024-1038.
- Wilson, D.J. and D.D.J. Netterville. 1976. Influence of downwind high-rise buildings on stack design. Jour. Air Pollut. Control Assoc. 26 (10): 976-980.
- Wong, R.K. 1985. A numerical model of small valley drainage winds and pollutant dispersion. Ph.D. Thesis, University of Alberta. 237 pp.
- Wong, R.K., K.D. Hage and L.D. Phillips. 1987. The numerical simulation of drainage winds in a small urban valley under conditions with supercritical Richardson numbers. J.Clim. Appl. Meteor. 26: 1447 - 1463.
- Woo, H.G.C., J.A. Peterka and J.E. Cermak. 1977. Wind tunnel measurements in wakes of structures. NASA contractor report CR-2806, March 1977.
- Yahalom, R. 1972. A preliminary study of the air flow and pollution in the Bow Valley region. Report No. 38, Department of Mechanical Engineering, University of Calgary. 34 pp.
- Yamada, T. 1981. A numerical simulation of nocturnal drainage flow. J. Meteorol. Soc. Japan. 59: 108-122.
- Yamada, T. 1979. An application of a three-dimensional simplified second-moment closure numerical model to study atmospheric effects of a large cooling-pond. Atmos. Environ. 13: 693 - 704.
- Yamada, T. and S. Bunker. 1988. Development of a nested grid, second moment turbulence closure model and application to the 1982 ASCOT Brush Creek data simulation. J. Appl. Meteor. 27: 562 - 578.

APPENDIX A

MEASUREMENT TECHNIQUES

A.1 MEASURING WIND

The increase of wind speed and the turning of wind direction with height poses a problem for making representative wind measurements, especially when comparisons are to be made. However, it turns out that wind direction changes in the lowest few tens of metres are negligible under homogeneous steady-state conditions although wind speed changes may be great.

For this reason the World Meteorological Organization (1972) has established that:

The standard exposure of wind instruments over level, open terrain is ten metres above the ground. Open terrain is defined as an area where the distance between the anemometer and any obstruction is at least ten times the height of the obstruction.

The wind so measured is called, by international convention, the surface wind. The word "surface" refers to the surface boundary layer where the frictional forces are nearly constant; it does not refer to the interface between the earth and the atmosphere where the wind is necessarily zero. The term "anemometer height" is sometimes used to denote the standard 10-m exposure.

Both wind speed and wind direction fluctuate in time. Therefore, an average over some time interval is implicit in any given value. In air pollution meteorology it is desirable to have winds averaged over intervals of about 1-h because this corresponds to the averaging times specified in most ambient air quality standards.

Wind averaging is not as straightforward as it may first appear. A true wind average would be a vector average, obtained by averaging individual components. The magnitude of this average vector would be the speed of the mean wind. However, the common cup anemometer actually measures the total horizontal component of wind speed, not just the speed along the mean wind direction. Averaging yields the mean speed of the horizontal component, not the speed of the mean wind vector. The former overestimates the latter by as much as 20% depending on the size of the wind direction fluctuations (Bernstein 1967). When there are large shifts in airflow as can occur in rough terrain, the vector mean speed may be essentially zero, even for a period without calms, whereas the cup anemometer would register some positive value.

Moreover, the response of a cup anemometer is described by a distance constant, which means that the response time varies with wind speed. Consequently the anemometer responds faster to increasing winds than to decreasing winds. This leads to an overestimation of the mean speed by 1 to 10% depending on the anemometer, measurement height and size of wind fluctuations (MacCready 1966, Hyson 1972).

Calm denotes the absence of measurable wind. Therefore, the occurrence of a calm is highly dependent upon the instrument used to measure the mean wind. The starting threshold is the wind speed at which the instrument begins to register speed values or direction changes. However, the response is asymmetric in that the instrument can register a wind below its starting threshold if the wind is decreasing from a higher value. "Dead" calms are virtually nonexistent. There is always some gentle motion of air and although this may be immeasurable with instruments, visual observations of smoke or other visible material will reveal the drift. Because pollution episodes are often associated with light winds, low-threshold instruments are desirable in air pollution meteorology. A number of such sensors are currently available with thresholds below 1 km/h and sufficient ruggedness for prolonged field use. A summary of available methods for measuring surface winds is given in Table A.1.

The largest source of long-term surface wind records in Alberta is the network of aviation weather stations operated by the federal Atmospheric Environment Service (Figure A.1). At most of these stations the standard wind equipment is the MSC type U2A which indicates the wind speed and direction on a pair of dials. The starting threshold of the system is about 1 m/s (2.1 knots for the cups and 1.4 knots for the vane) and the resolution is about 0.5 m/s (1 knot) for speed and 10° for direction.

The wind speed and direction reported by these stations are visual averages of the readings on the dials over a 1-minute interval (2 minutes from 1985 to present) close to the hour. The wind speed is recorded to the nearest knot and the wind direction to the nearest 10 degrees. Western Research & Development (1976) found that for low wind speeds associated with urban air pollution events, such observations tend to be systematically higher than corresponding 1-h averages from a chart recorder. This tendency may reflect a human observer "set" due to the concern of the aviation industry for high winds, or it may represent an actual sampling error such that a 1-minute average

Table A.1 Anemometers suitable for diffusion applications.

Anemometer Type	Principle of Operation	Response	Notes
rotating cup	aerodynamic forces using kinetic energy of wind	slow	senses total horizontal wind
rotating propeller		slow to moderate	correction required for lack of cosine response
pressure sphere		fast	not commercially available; extensive signal processing required
drag sphere		fast	marginal threshold
hot wire	thermodynamic forces using cooling power of the wind	very fast	too delicate for routine use
hot film		very fast	no moving parts
sonic	acoustic properties using effect of wind on wave propagation	very fast	expensive
vortex		fast	marginal threshold

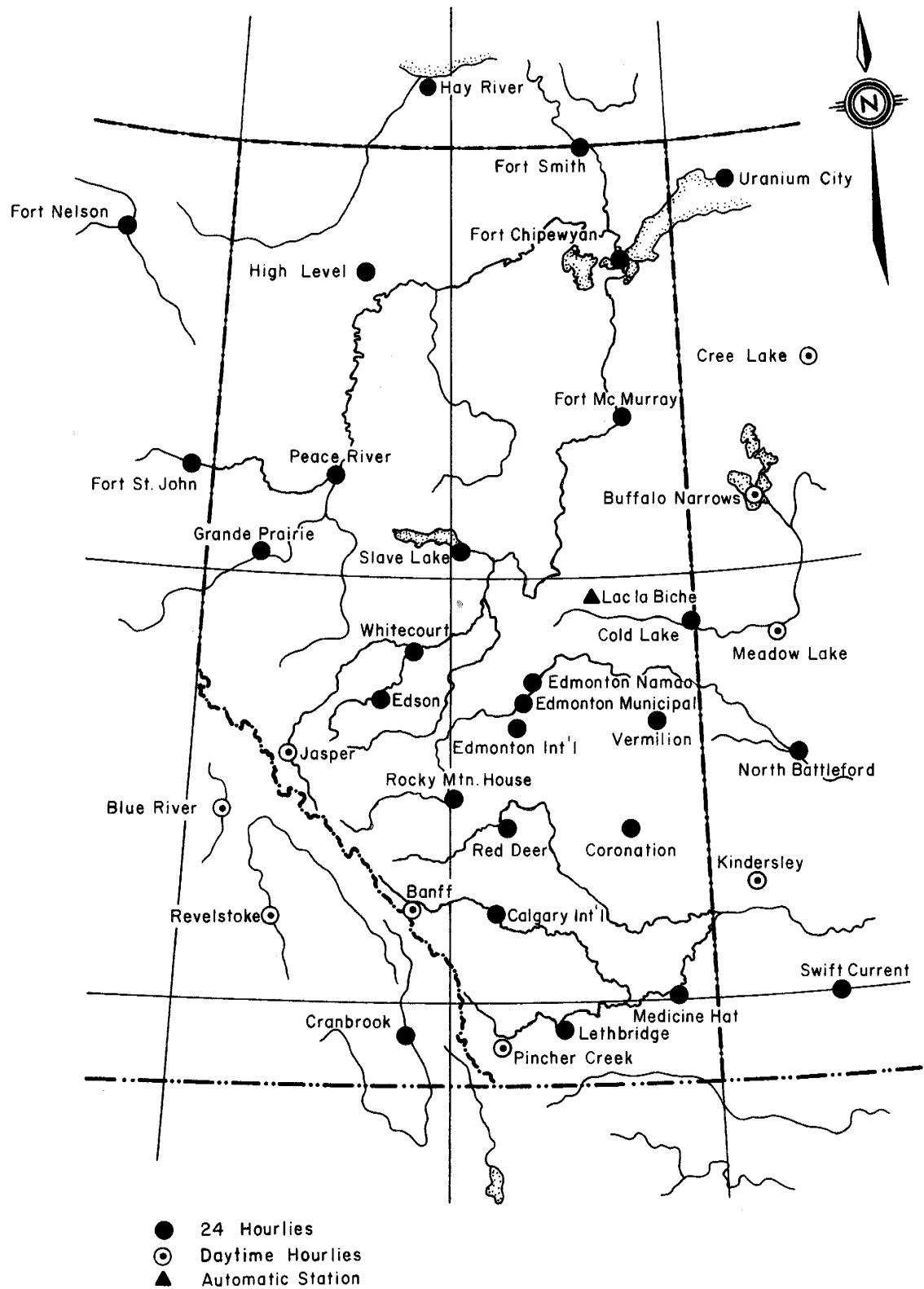


Figure A.1 Weather stations operated by the Atmospheric Environment Service in and near Alberta.

is not an unbiased estimator of the 1-h average, at least for light winds. It has been shown that even 2-minute wind averages are marginally useful for representative wind determination (Wieringa 1980). The 10-minute average used for synoptic observations was found to be much more representative of the hourly average.

Environment Canada (1982) publishes tables showing the average percent direction frequency for each month and the whole year, together with the mean speed for each direction and for all directions. In addition, the height of the anemometer at each station and a few brief notes on exposure are provided. The latter are not always accurate or up-to-date. Some stations do not meet the WMO criteria for height and exposure. The serious user of aviation wind data ought to visit the station in person and obtain the station history from the Regional Office of the Atmospheric Environment Service in Edmonton.

Over the years the number of air quality monitoring stations in the province has grown steadily and, since wind is generally one of the parameters being measured, an extensive data bank is gradually being assembled. Most such stations are operated by industries in accordance with the monitoring provisions of a Licence-to-Operate under the Alberta Clean Air Act. The choice of site and instrumentation must conform with specifications set out in the Air Monitoring Directive (Alberta Environment 1989). Wind instruments are required to have a starting threshold no greater than 1.5 km/h. Exposure and height must meet the WMO standard or definite alternative criteria derived from aerodynamic considerations. Complete site documentation, including photographs, is also required. The wind speeds reported are hourly averages to the nearest kilometre per hour; wind directions are hourly averages in terms of an eight-point compass, if done by eye from analog strip charts; or to the nearest degree if performed by microprocessor systems.

Unfortunately these data are not easily accessible, nor are summaries readily available. The annual air monitoring report from each industry to Alberta Environment usually contains a summary of the data for that year. Sometimes a 5-year summary is provided as part of the site documentation. For information about such data the potential user should contact the industrial operator of the station.

Above about 100 m it is impracticable to use cup anemometers or indeed any tower mounted instrument for measuring wind speed and direction. The winds aloft are usually determined by tracking the motion of an expandable balloon inflated with hydrogen or helium to rise at a nearly uniform rate. By locating the balloon in space at given intervals, the displacement vector can be determined. Dividing the magnitude of the displacement vector by the time interval yields the mean speed of the wind for the layer traversed by the balloon. There are two problems associated with this technique:

1. The balloon does not sample the wind at each level at the same time. This produces an error which depends on the nature of the vertical profile of the wind fluctuations and the response of the balloon.
2. The balloon does not sample the wind directly above the release point. The sampling points move progressively farther away with time. This means that it is impossible to determine the vertical variation of wind above a fixed point on the earth's surface.

It is tacitly assumed that a balloon is immediately accelerated to the velocity of the wind at any given level because its surface area is high and its inertia is low, but this does not seem to be the case for radiosonde balloons (Scoggins 1965). Also, rising balloons exhibit self-induced lateral motions that can lead to significant errors unless relatively thick layer averages are considered (Murrow and Henry 1965). Smooth balloons tend to move in a regular spiral or zigzag path, but representative layer winds are obtained provided the layer considered has a thickness equal to several wavelengths of the spiral motion or greater than 15 balloon diameters (MacCready 1965).

In addition to the foregoing problems with the technique there are several sources of observational error, depending upon the tracking method used. For altitudes of up to 5 km, methods that measure slant range have typical vector errors of 0.5 m/s to 1 m/s while methods that measure only angles have typical vector errors of 0.5 m/s to 4 m/s depending on the ratio of mean wind to ascent rate (WMO 1971). Even the best observations are only within $\pm 2^\circ$ for direction and ± 1 m/s for speed at low altitudes, while at 5 km these inaccuracies may be doubled (Middleton and Spilhouse 1953). In a comparison of tower and balloon measurements, there were mean differences of 15° in

direction and 1.9 m/s in speed at a height of 150 m (Rider and Armendariz 1966). Netterville and Djurfors (1979) have shown that the errors grow quickly with elapsed time of balloon flight and that progressively longer observation intervals are required to maintain a constant percentage error at greater altitudes.

Winds aloft are determined twice daily at each of the Upper Air Stations operated by the Atmospheric Environment Service (Figure A.2). Observation times are standardized world-wide as 00:00 Z and 12:00 Z (Coordinated Universal Time) but the actual time of release of the balloon-borne radiosonde is 45 minutes earlier. For Alberta's only upper air station – Edmonton (Stony Plain) – this means release times of 16:15 MST and 04:15 MST. Radio theodolites automatically track the radiosonde transmitter and record the azimuth and elevation angles from the ground-based receiving station. Coupled with the known pressure height from the radiosonde aneroid the position of the package can be determined and the mean wind calculated. A number of aviation weather stations release pilot balloons to determine winds aloft, but these data are not readily available for climatological purposes.

Remote sensing techniques have developed rapidly in the past two decades and can now be used for routine upper air wind measurements. Remote sensors are classified as monostatic if transmitter and receiver are located in the same spot or bistatic if they are separated by a substantial baseline but aimed at the same volume in space. Remote sensors provide spatial averages over volumes of air rather than the single-point measurements given by in-situ sensors like anemometers. Radio, light or sound waves may be used, but SODAR (sound detection and ranging) has found much greater application than other methods in air quality work, largely because of its lower cost. Davies (1982a) found that Doppler SODAR wind measurements compared favorably with winds measured by pilot balloon in northeastern Alberta. A summary of remote sensors is given in Table A.2.

A number of SODAR units have been operated by the University of Calgary, Physics Department; analysis of early data can be found in Mathews and Hicks (1984). Suncor has collected data in the Athabasca oil sands area near Fort McMurray as have Syncrude and Canstar (Davies 1982b). Husky operates a unit at their Ram River Gas Plant near Rocky Mountain House. Alberta Environment collected SODAR data southwest of Edmonton from 1982 to 1987. The first year's data were analyzed by Leahey

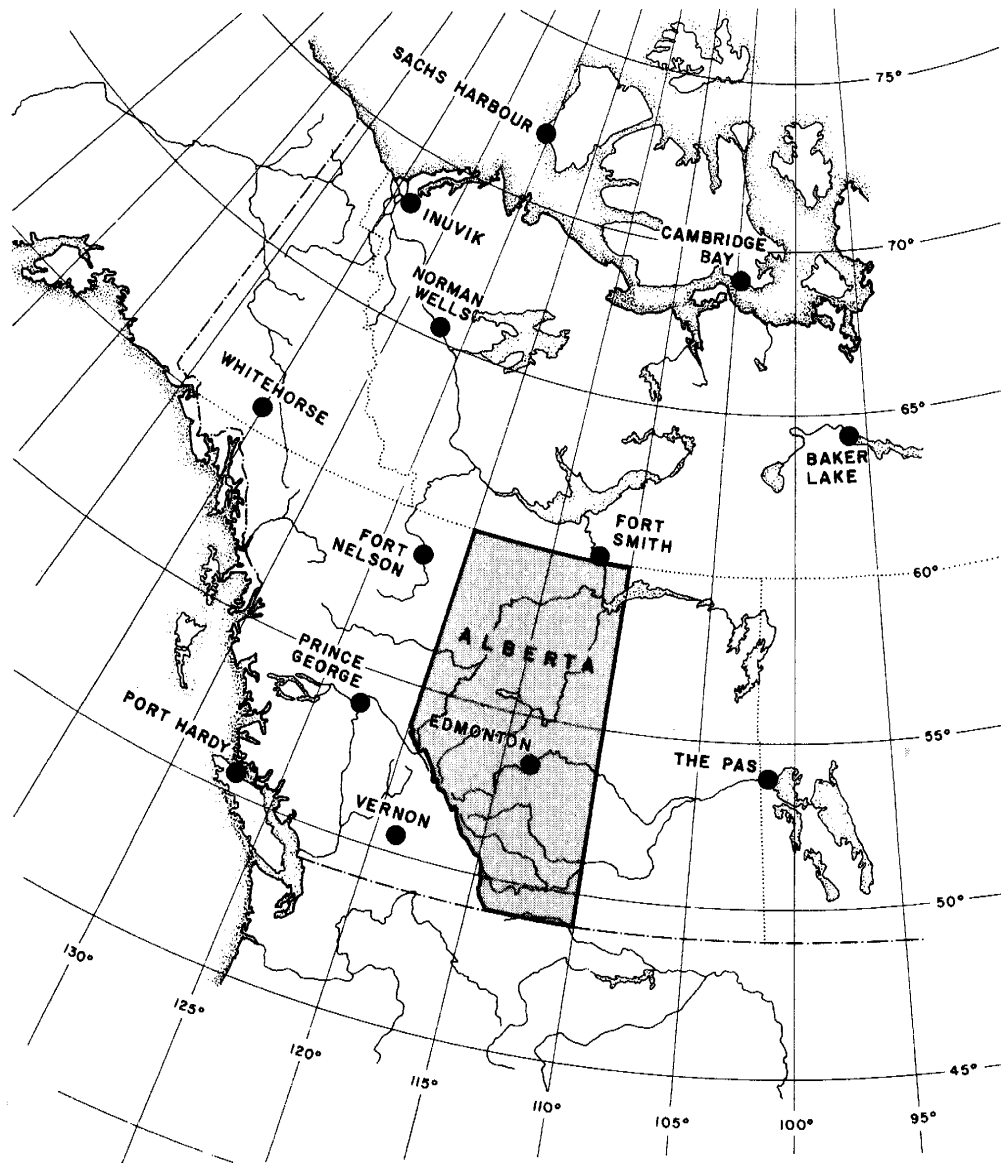


Figure A.2 Upper air stations operated by the Atmospheric Environment Service in western and northwestern Canada.

Table A.2 Remote sensors for winds above the surface layer
(after Schwiesow 1986).

	SODAR (sound detection and ranging)	LIDAR (light detection and ranging)	RADAR (radio detection and ranging)
Minimum Altitude (m)	10	20	30
Maximum Altitude (m)	1500	10,000	50,000
Altitude Resolution (m)	20-50	5-300	25-300
Accuracy (m/s)	0.2-0.4	0.2-2	0.2-2
Disadvantages	. sensitive to external noise . return success is variable and weather-dependent	. possible danger to eyes . beam attenuated by cloud and fog	. large and expensive . requires high power . side lobes limit use near ground

and Hansen (1985a, b). Winds aloft were also measured at this site by pilot balloon five times per day during mid-season months from 1977 to 1983. A scaled down version of this program operated in the Fort Saskatchewan area as a joint venture between the Fort Saskatchewan Regional Industrial Association and Alberta Environment from 1988 to 1991.

A.2 MEASURING TEMPERATURE STRATIFICATION

A variety of in-situ and remote sensing methods are available for measuring temperature as a function of height above ground (Table A.3). Only the first three listed in the table have been used in Alberta.

Rawinsonde data have several inherent limitations. The temperature sensors employed have a nominal accuracy of $\pm 0.4^{\circ}\text{C}$. The pressure-dependent cycling of information channels results in a vertical resolution of about 125 m. Heights can be calculated to within ± 30 m but data obtained in the first 100 m above the ground are unreliable.

The main error associated with the minisonde lies in the determination of height above ground. The rise rate of the 30 g (6 in.) lift balloon is nominally 3 m/s but Longley (1956) reported increases of up to 25%, and standard deviations of about 0.4 m/s have been reported. The standard practice of the Atmospheric Environment Service is to allow an increase of 20% in rise rate for the first minute, 10% for the second and third minutes and 5% for the fourth and fifth minutes. Angle and Godin (1979) reconfirmed this result with double theodolite observations taken at several Alberta locations. The use of two tracking theodolites eliminates the rate-of-rise assumption and for typical operating conditions leads to errors in balloon height of 1 to 2% and uncertainties in the gradient of 10 to 20%. Netterville and Djurfors (1979) have shown how these errors grow with increasing time from release, decreasing length of baseline and decreasing observation interval. Choukalos (1977, 1978) gives a computer program for processing minisonde data using the method of Thyer (1962).

Long-term data on the thermal stratification of the entire troposphere are available from the Canadian Upper Air Network, which has but one station in Alberta, at Edmonton (Stony Plain). Reporting is by means of significant levels which are selected such that the temperature profile is reproduced to within 1°C of the actual observed

Table A.3 Methods for measuring temperature stratification in the atmospheric boundary layer (after Barnes 1972 and Schwiesow 1986).

Method	Accuracy of Temperature (°C)	Altitude Resolution (m)	Limitations
Instrumented Tower	<0.2	<1	< 300 m; most installations are about 100 m high
Rawinsonde	0.3 - 0.8	125	Lowest 100 m unreliable
Minisonde	0.1 - 0.2	5	Rise rate uncertain unless tracked by two theodolites
Aircraft	0.5	8	Expensive
Drones Dropsonde Rockets	0.5	8	Recovery can be difficult; dangerous in built-up areas
Passive Microwave	1-4	15	Cannot operate in rain or clouds
Lidar	1	5-150	Danger to eyes; not all-weather
Radio Acoustic Sounding (RASS)	0.5	10	Little operational experience

values. The soundings are made twice daily, nominally at 05:00 MST and 17:00 MST irrespective of season. (The actual release of the balloon usually occurs 45 minutes prior to these times). Although the measurements are rather coarse, they can provide a general picture of long-term average temperature stratification.

Two instrumented 100-m towers were in operation in Alberta from 1974 to 1980, one in Calgary and one in Edmonton, as documented by Choukalos (1985). Ten-minute averages beginning on the hour are extracted and tabulated. This gives a reasonably good picture of changes in time, although it is necessarily restricted to a very shallow layer near the ground.

Monostatic acoustic radar or SODAR provides a continuous picture of mechanical mixing height in a stable layer. Mathews and Hicks (1978) operated such a device in Calgary from March 16, 1976 to July 31, 1977. A similar unit was operated by Angle and Godin (1982) near Edmonton from 1977 to 1982.

A mini-sounding program, described by Angle and Godin (1979) and a 20-m tower, documented by Sakiyama and Wong (1987), also operated near Edmonton from 1977 to 1982. Murray (1977, 1983) and Murray and Kurtz (1976) reported on a number of soundings in northeastern Alberta near the Syncrude plant. James F. MacLaren Limited (1978) summarized data collected by the Alberta Oil Sands Environmental Research Program. A number of towers and sounding programs have been operated by industry in connection with Environmental Impact Assessments, but the data are not published and are not readily accessible.

A.3 TURBULENCE

Atmospheric turbulence can be measured by any of the fast-response methods used to measure the mean wind (Kaimal 1981). Combined with measurements of temperature stratification, values for the gradient Richardson number can be calculated.

The common cup anemometer is not well suited to turbulence measurements because (1) it measures the horizontal winds $u_h = (u^2 + v^2)^{1/2}$ rather than an individual component (Bernstein 1967), (2) it cannot measure turbulent energy at higher frequencies (MacCready and Jex 1964a). If the cups have a distance constant of less than 1 m it is

possible to get an estimate of the dissipation rate (Frenzen 1965). The high-frequency cutoff is a serious limitation only when turbulence levels are low or the small scales are of interest. The large energy-containing eddies are of paramount importance in structural engineering. The standard deviation of the horizontal wind speed is a useful measure of "gustiness" for these applications.

The standard deviations of the three wind components should ostensibly be measurable by means of an array of sensors mounted at 90° to one another so that each sensor responds to one wind component. The three time series generated could then be subjected to the various analyses discussed in the previous section. Unfortunately one cannot know in advance what the mean wind direction will be, so it is impossible to orient the array so that one sensor is responding to the longitudinal component (u) only. Consequently the "u and v" time series must be stored, the mean wind direction computed, and then the two actual wind components determined with respect to this rotated coordinate system T. A microcomputer can do this on site or the data can be logged and a central computer used to process the information after the fact (Horst 1972, 1973). It is important to realize that the so-called "u-v-w orthogonal arrays" do not give the u and v components directly. Such sensor arrays can be made up of either sonic anemometers or helicoid propeller anemometers (Hicks 1972, Kaimal 1975, Gill 1975). Propeller anemometers also require corrections to compensate for this lack of cosine response (Bowen and Teunissen 1986, 1984; Teunissen et al. 1985).

The vertical component of turbulence was measured at Alberta Environment's Ellerslie site for a period of 5 years using a Gill propeller anemometer. Nighttime values were typically 0.15 m/s irrespective of season, but daytime values ranged from 0.20 m/s in the colder months to 0.46 m/s in April and May (Sakiyama and Wong 1987).

In general the instantaneous wind vector will lie at an angle e (for elevation) from the horizontal and at an angle a (for azimuth) from the position of the mean wind vector. For small angles and longitudinal turbulence intensities that are not too large it can be shown that:

$$i_y = \frac{\sigma_v}{u_{avg}} \approx \sqrt{(a^2)_{avg}} = \sigma_a$$

$$i_z = \frac{\sigma_w}{u_{avg}} \approx \sqrt{(e^2)_{avg}} = \sigma_e$$

Therefore the fluctuations of a vane that is free to pivot in both the horizontal and vertical planes can be used to measure turbulence intensities. Such a device is known as *bivane* (bidirectional vane) and has the advantage that the signals may be processed electronically by means of a *sigma meter*.

Digitally recorded bivane data were used by Walker (1964) to construct turbulent intensity spectra for the rolling prairie near Suffield (Figure 1.29). Leahey (1974) and Leahey and Davies (1978) used bivanes at several Alberta sites. Sakiyama and Wong (1987) reported on bivane measurements at Ellerslie: lateral intensities ranged from 5% in winter to 31% on spring afternoons. Vertical intensities ranged from 4% to 24%.

It follows that the wind vane used to determine mean wind direction can also be used to measure cross-wind turbulence intensity. It has been recommended that sigma meters be installed on the wind direction vanes at all U.S. National Weather Service observing stations (Carlson et al. 1978). Hourly σ_a was determined at Alberta Environment's Ellerslie station for 7 years. In winter, averages ranged from 0.7° at night to 15° in the afternoon. In summer, averages ranged from 11° at night to 29° in the afternoon.

Numerical values for any of the measures of turbulence depend upon sensor and sampling characteristics. The interval during which data are recorded for determination of the mean is called the *averaging time*, the *sampling duration* or the *length of record*. Instantaneous values may be subjected to a moving average that smooths out the smallest fluctuations. The interval used for this pre-averaging is called the *smoothing time* (sometimes it is called the averaging time, and this can lead to considerable confusion). The inertia of a mechanical wind sensing system will automatically impose a smoothing due to the inability of the system to respond faithfully to very high frequency fluctuations. If the averaging time is represented by T , the smoothing time by s , and the instrument response to a step change is an exponential with time constant α , then eddies with periods between T and $10s$ (or 30α) are satisfactorily included but those with periods greater than $10T$ and less than s (or α) are excluded (Pasquill 1974). In other words both the low- and high-frequency ends of the spectrum are cut off. The variance of the sample is, therefore, generally less than the variance that

would have been obtained using instantaneous values (infinitesimal smoothing) and infinite sampling duration. The application of a smoothing and sampling process to a time series constitutes a filter or a "window" through which only a portion of the total energy in the spectrum can be observed.

In practice, the variance of the elevation and azimuth angles may increase with sampling time only up to some point and thereafter remain constant. This will occur if there is a spectral "gap," that is, little energy in eddies having comparable periods. This condition is often satisfied for the elevation angle but rarely for the azimuth angle. If the sampling time is increased beyond the gap then the variance again increases as a new energy-containing region of the spectrum is sampled.

A vane is characterized by a natural wavelength related to the "stiffness" of the rotation and a damping ratio which describes the ability to track direction changes accurately. The most faithful response of a vane is obtained when the damping ratio (ratio of actual damping to critical no-overshoot damping) is about 0.6 which corresponds to a 10% overshoot for step disturbances (Moses 1968). Vanes with damping ratios less than 0.4 require the removal of resonant oscillations by smoothing over 10- or 20-s intervals.

The indicated variance from a direction vane with a damping ratio of about 0.43 is the same as for a vane with perfect response (MacCready and Jex 1964b). For this damping ratio the overshoot near the natural wavelength compensates for the loss by vane attenuation at shorter wavelengths. If the natural wavelength of the vane is very short (less than 0.4 times the height of measurement) the exact value of the sampling ratio is unimportant. This result is only valid for the inertial subrange. Since the largest eddies in the inertial subrange are roughly equal to the height of measurement, this implies a short sampling duration of $T = z/u$.

A simple measure of variability is the range of values in a given sample. The wind direction range is well correlated with the azimuth angle standard deviation for all smoothing times and sampling durations (Markee 1963). For smoothing times of 1 to 10 s the range divided by 6 gives a rough estimate of the standard deviation, that is, $\sigma_a = R/6$. Alternatively it has been suggested that a few of the largest deviations from the mean be discarded and the range be divided by 4.

The output from a sensor is generally a continuous current or voltage varying in proportion to the parameter being measured, and is known as an *analog signal*. For computation purposes this signal is often converted to a discrete series of numbers representing the amplitude of the signal at different instants of time. This process is called digitization and the period of time between values in the number series is called the *digitization interval*. (To add to the confusion, this time period is also called the sampling interval). The digitization rate or sampling frequency is the number of discrete values obtained in unit time.

The ratio R/σ for angles measured by a bivane over 1-h periods decreases with digitization interval and the relation $\sigma = R/6$ is valid only for an azimuth digitization interval of 15 s and an elevation digitization interval slightly larger than 60 s (Pendergast and Crawford 1974). Range-standard deviation relationships are sensitive to recorder response, extraneous signals, data reduction procedures and the scale of turbulence (Mooney and Swanson 1976). To minimize these, the range is determined for the last 10 minutes of the hour after excluding both the extreme maximum and extreme minimum excursions. (The removal of extremes reduces the effects of spurious signals and the 10-minute sampling time approximates the lower frequency limit of "micrometeorological turbulence"). With this type of system $R/\sigma = 4$ for azimuth angle and $R/\sigma = 5$ for elevation angle; 90% of all observations are within a factor of 2. Stable conditions produce significantly lower values of R/σ .

The calculation of the standard deviation of azimuth angle fluctuations is by no means trivial because wind direction is a circular variable with a crossover between 360° and 0° . If all instantaneous values over the averaging period are retained until the end, then a straightforward "two-pass" method can be used to account for the crossover. The resultant direction is determined and subtracted from 180° . This adjustment is then added to each individual value and the sum of squares calculated (Turner 1986) for substitution into the standard statistical formulas. Since this procedure requires a large amount of storage for each individual value, a variety of "single-pass" shortcuts have been developed:

- (1) analog electrical determination (Jones and Pasquill 1959, Brock and Provine 1962), the original *sigma meter*,

- (2) accumulation of unidirectional angles and number of direction reversals (Harris and McCormick 1963),
- (3) averaging swing amplitudes (Turner 1968),
- (4) accumulation of sines and cosines (Verrall and Williams 1982),
- (5) component statistics (Ackermann 1983),
- (6) sine and cosine estimators (Yamartino 1984),
- (7) circular variance for a wrapped normal distribution (Mardia in Fisher 1983 and Mori 1986),
- (8) short-term wind rose (Nelson 1984),
- (9) incrementing the previously sampled direction (Mori 1987),
- (10) vector and scalar mean winds (Yamartino 1984, Mori 1986),
 $V_v/V_s = \exp(0.5 \sigma_d^2)$ for $\sigma_d < 20^\circ$.

Many of these methods work well for small angular fluctuations but fail when large fluctuations are encountered. Part of the difficulty stems from the use of three different definitions of mean wind direction: (1) The direction of the vector mean wind, (2) the arithmetic mean wind, (3) the direction of the mean of the unit vectors (Mori 1986). Another part of the problem stems from the use of linear statistics for a circular variable (Fisher 1987).

The difference between vector mean wind spreads (V_v) and scalar mean wind speeds (V_s) can also be used to estimate the standard direction of the azimuth angle fluctuations (Yamartino 1984) or to calculate an exact value for the cross-wind component from the equation (Mori 1986)

$$\sigma_v = [2 V_v (V_s - V_v)]^{1/2}$$

The ability of SODAR to measure wind components in layers up to several hundred metres means that turbulence profiles can also be determined. However, since new values can be obtained only after a few minutes (during which the sound spectra are accumulated to determine the Doppler shift), this method is limited to the larger scales of turbulence. It has been applied in the Fort McMurray area by Netterville (1981) and Davies (1982a, b) and in the Edmonton area as reported by Leahey and Hansen (1985). Balser and Netterville (1981) showed that the variance of the Doppler spectra can be used to estimate the contribution of the smaller scales. Bara and Netterville (1985) showed how to derive Eulerian integral length scales from SODAR data and reported a value of 43 m for lateral turbulence at 200 m above ground during mid-morning in early April.

Instrumented aircraft have two advantages over ground-based measurement techniques:

- (1) They can obtain larger samples in a given time period because the platform is moving at many times the wind speed;
- (2) They can obtain samples representative of a large area rather than a single point.

A number of turbulence flights have been made in northeastern Alberta by Davison et al. (1976) and Davison and Grandia (1979). A sample of the results is presented in Table A.4.

The fluxes of heat and momentum, although fundamental to stability considerations, are among the most difficult quantities to measure. Four main methods have been developed:

- (1) Eddy correlation methods (Hicks 1970, Kaimal 1969) require fast response sensors (McBean 1972), matched output and large storage capacity to allow the calculation of covariances either on-line or after the fact (Kaimal 1975). Serious errors can arise if the vertical component is not aligned correctly (Kaimal and Haugen 1969) and in general only about 20% accuracy is anticipated (Kaimal 1986).

Table A.4 Summary of turbulence data from aircraft flights in the Fort McMurray area.

(a) March 11, 1976 12:50 - 16:55 MST (Davison et al. 1976)

Height (m MSL)	σ_w (m/s)	σ_u (m/s)	σ_v (m/s)	$(w'u')_{avg}$ (m/s) ²	$(w'v')_{avg}$ (m/s) ²	$(w'T')_{avg}$ (°C m/s)
1370	0.54	0.9	*	0.11	*	0.018
1220	0.65 (0.10)	1.32 (0.13)	*	0.05 (0.09)	*	-0.040 (0.03)
1070	0.57	1.4	*	0.43	*	0.012
910	1.21	1.8	2.1	-0.27	0.10	0.030
760	1.26	2.2	*	-1.28	*	0.033
610	1.10 (0.03)	1.66 (0.20)	2.8	-0.36 (0.07)	-0.85	0.000 (0.006)
550	1.17	2.3	*	-0.21	*	-0.033
460	1.07	2.8	*	0.24	*	0.030

(b) 19 June 1977, 07:45 - 10:10 MDT (Davison and Grandia 1979)

Description	σ_w (m/s)	σ_{uH} (m/s)	ϵ (cm ² /s ³)	L_w (m)	L_v (m)
Early Above inversion	0.99 (0.05)	0.93 (0.08)	5 (1)	260 (27)	240 (36)
Early, Below inversion	1.30 (0.08)	1.10 (0.10)	29 (3)	195 (17)	125 (32)
Later Below inversion	1.55 (0.10)	1.81 (0.18)	55 (7)	200 (14)	230 (34)

Note: Values in parentheses are the standard deviations of the population of mean values.

σ_w	standard deviation of vertical velocity
σ_{uH}	standard deviation of lateral velocity with respect to the aircraft
ϵ	dissipation
L_w	integral length scale in the vertical
L_v	integral length scale in the lateral direction with respect to the aircraft

- (2) Profile methods (Wagonner 1965, Panofsky 1963, Panofsky and Dutton 1984) require at least two levels of mean wind and temperature measurements. The profile equations are then solved interactively to obtain the heat and momentum fluxes (and the roughness length if three or more levels are available). The measurement site must be surrounded by a uniform surface for a distance at least 100 times the height of the highest measurement to ensure that the profiles are in equilibrium with the underlying surface fluxes as noted by Angle (1975).
- (3) Variance methods (Hicks and Wesley 1978, Eilts et al. 1987) rely on the observed correlations between turbulent fluctuations, which can be written as $F_p = R_{wp} \sigma_w \sigma_p$ where σ_w is the vertical wind standard deviation, σ_p is the standard deviation of any parameter of interest, and R_{wp} is the correlation coefficient between the two which has a value of about 0.4.
- (4) Eddy accumulation methods (Desjardin 1977) use a valve to collect separate air samples for updrafts and downdrafts. The flux of any scalar is then calculated from the differences between the means in the two containers.

Momentum flux can also be determined from lysimeter-like drag plates (Bradley 1972), from drag coefficients once surface properties are known (Bradley 1972) or from bivariate elevation angle offsets (Chimonas 1980). Values of the fluxes can also be inferred from autocorrelation of the vertical wind (Misra 1979) and the variance of difference between heights (Wyngaard and Clifford 1978).

The surface heat flux is also the major driving mechanism of mixed-layer growth. Both the height of the convective boundary layer and the nature of the turbulence therein are controlled by the heat flux. Despite its importance it is not routinely measured, largely because of the stringent instrument requirements. Even the most sophisticated sonic anemometer/thermometer eddy correlation system can probably do no better than 20% accuracy, plus it is very difficult to ensure areal representativeness. Estimation methods using more readily available measurements have been discussed in Chapters 1 and 2.

A.4 BOUNDARY LAYER HEIGHT

A number of possible definitions of the thickness of the atmospheric boundary layer were mentioned in Chapter 2, all based upon profile of wind, temperature and turbulence. Remote sensing techniques introduce a few more alternatives, defined in terms of the physical quantity that is being measured. There are summarized in Table A.5.

A.5 DILUTION

All of the prediction schemes discussed in Chapter 5 were based on field studies involving measurement of:

- (1) The ambient concentrations of a pollutant or deliberately cleared tracer material.
- (2) The rate of release of the pollutant or tracer material.
- (3) The concurrent meteorological conditions.

Descriptions of these types of field studies appear in the literature from the 1950s when smoke, lycopodium spores and fluorescent particles were used, to the present where a variety of perfluorocarbons and other gaseous tracers are used in conjunction with advanced analytical capabilities. Reviews of tracer technology (Deuble 1979, Dietz and Senum 1986) and existing data bases (Woodruff and Glantz 1985, Vanderborght and Kretzchmar 1984) are available.

Direct dilution measurements are useful not only for research purposes, but also to confirm model predictions or to quantify the dilution rates in very complex situations. For the short-range diffusion of primary concern with industrial sources, the most cost-effective tracer is sulphur hexafluoride. Details of study design are well documented (Johnson 1983, 1985; Sackinger et al. 1982). A summary of information about SF_6 tracer techniques is given in Table A.6. Two such studies have taken place in Alberta: Moose Mountain, southwest of Calgary reported by Sakiyama and Angle (1988) and Sakiyama (1983); and Kathryn, northwest of Calgary reported by Holizki and Strom (1989).

Table A.5 Methods for measuring the thickness of the atmospheric boundary layer
(after Coulter 1979 and Kaimal et al. 1982).

Method	Physical Principle for Definition
LIDAR	sudden decrease in detectable particle concentration
SODAR	sudden change in small-scale temperature differences
balloon-borne sonde	(1) first kink in the temperature profile (2) maximum in wind profile (3) wind direction first becomes geostrophic
RADAR X-based FM-CW TPQ-11	drop in echo intensity from artificial chaff discontinuity in refractive index sudden decrease in particulates

Near-source plume behavior can often be studied photographically if the plume can be made visible. Davies (1986) reviewed available plume visualization techniques and concluded that ultraviolet methods show the most promise as they are not dependent on injections of foreign substances or specific atmospheric conditions.

Table A.6 Equipment and design characteristics for SF₆ tracer studies.Fixed Samplers

Sampling methods and costs: tedlar bags (\$1-5), plastic syringes (\$1)

Analysis equipment: gas chromatograph with electron capture detector (\$15,000)

Analysis time: < 10 min per sample

Detection limit: 1.0×10^{-6} ppm

Sampler Layout		
Downwind distance (km)	1.0	10.0
1 h plume σ_y (F stability, 10 cm roughness)	62	470
1 h plume width, $4\sigma_y$, (m)(°)	248(13.9)	1880(10.6)
maximum sampler separation* (m)(°)	50(2.8)	380(2.1)
number of samplers for 120° arc	43	57

Mobile Analyzer

Analysis equipment: continuous flow sealed electron capture detector (\$23,000)

Detection limit: 10.0×10^{-6} ppm

Response time: < 15 s

Sampling method: right-angle transect across plume at fixed speed

Maximum transect speed: 10 km/h at 1 km, 70 km/h at 10 km

Number of transects needed per atmospheric condition**: 6-24

Sulphur Hexafluoride

Ambient background: $< 0.5 \times 10^{-6}$ ppm

Cost: \$15/kg

Typical release rate: ground source 0.1 g/s
elevated source 1 g/s

* At least five measurements must be taken in the plume ($\pm 2\sigma_y$) to resolve position and spread within 15% (Sackinger et al. 1982; Johnson 1983, 1985).

** The smaller number will generally yield values within $\pm 40\%$ of the true values with 95% confidence; the larger number will yield values within $\pm 20\%$ of the true values with 95% confidence.

A.6 REFERENCES

- Ackermann, G. R. 1983. Means and standard deviations of horizontal wind components. *J. Clim. Appl. Meteor.* 22: 959 - 961.
- Alberta Environment. 1989. Air Monitoring Directive (AMD). Monitoring and Reporting Procedures for Industry. Environmental Protection Services. Edmonton. 376 pp.
- Angle, R.P. and J.U. Godin. 1982. Toward an atmospheric boundary layer climatology for Central Alberta. Current climatological activity in Alberta. Proc. of the 6th Annual Meeting, Alberta Clim. Assoc., 25 Feb. 1982. 25-37.
- Angle, R. P. and J. U. Godin, 1979: A meteorological station for air quality management. Paper 79-10.2. 1979. Annual Meeting of Pacific Northwest International Section - Air Pollution Control Association, November 7 - 9, 1979.
- Angle, R.P. 1975. Airflow from mustard to fallow. *Atmosphere* 13: 110-125.
- Balser, M. and D. Netterville. 1981. Measuring wind turbulence with Doppler - acoustic radar. *J. Appl. Meteor.* 20: 27 - 35.
- Bara, B. and D.D. J. Netterville. 1985. Determination of plume "sigmas" and turbulence integral length scales from acoustic Doppler measurements of plume-level wind fluctuations. PNWIS/APCA Joint Annual General Meeting, Calgary, November 13 - 15, 1985. 503-571.
- Barnes, A.A. 1972. Temperature sounding up to 2 km. A review of techniques. 2nd Symp. on Meteor. Obs. and Instr., San Diego, March 27 - 30. Amer. Meteor. Soc., Boston.
- Bernstein, A. B. 1967. A note on the use of cup anemometers in wind profile experiments. *J. Appl. Meteor.* 6: 280 - 286.
- Bowen, A. J. and H. W. Teunissen. 1986. Correction factors for the directional response of Gill propeller anemometers. *Boundary Layer Meteor.* 37: 407 - 413.
- Bowen, A. J. and H. W. Teunissen. 1984. Correction factors for the directional response of propeller anemometers, Research Report No. MSRB-84-2, Atmospheric Environment Service, Downsview, Ontario, Canada.
- Bradley, E. F. 1972. The influence of thermal stability on a drag coefficient measured close to the ground. *Agric. Meteor.* 9: 183 - 190.
- Brock, F.V. and D.J. Provine. 1962. A standard deviation computer. *J. Appl. Meteor.* 1: 81 - 90.
- Carlson, J.H., R.A. Hollister, M. E. Smith and F. P. Castelli. 1978. Inadequacies of atmospheric stability measurements and recommendations for improvement. pp. 213 - 223. *Air Quality Meteorology and Atmospheric Ozone*, ASTM STP 653, A. L. Morris and R. C. Barras (ed.), Amer. Soc. for Testing and Materials 1978.

- Chimonas, G. 1980. Reynolds stress deflections of the bivane anemometer. *J. Appl. Meteor.* 19: 329 - 333.
- Choukalos, M. 1985. Summary of data from Tall Meteorological Towers in Calgary and Edmonton. Alberta Environment. March 1985. 89 pp.
- Choukalos, M. 1978. A computer program for processing data from single or double theodolite tracking of minisondes. Volume 2. Processing extension and subroutine for analysis and plotting. Alberta Environment.
- Choukalos, M. 1977. A computer program for processing data from single or double theodolite tracking of minisondes. Alberta Environment. August 1977.
- Coulter, R.L. 1979. A comparison of three methods for measuring mixing-layer height. *J. Appl. Meteor.* 18: 1495 - 1499.
- Davies, M.J.E. 1986. Plume visualization techniques. Prep. for Standards and Approvals Division, Alberta Environment by Concord Scientific Corporation. 27 pp.
- Davies, M.J.E. 1982a. An intercomparison of wind profiles obtained from a Doppler Acoustic Wind Sensor with those obtained from tracking helium filled balloons. Presented at the Pacific Northwest International Section of the Air Pollution Control Association Meeting, November 15-17, 1982, Vancouver, British Columbia. 2B3.1-2B-3.6.
- Davies, M.J.E. 1982b. The use of a Doppler acoustic wind sensor to obtain upper level wind data for plume dispersion calculations. Environmental Symposium, Ancon, Ecuador, 9 - 12 May 1982. 23 pp.
- Davison, D.S. and K.L. Grandia. 1979. Plume dispersion measurements from an oil sands extraction plant, June 1977. prep. for the Alberta Oil Sands Environmental Research Program by Intera Environmental Consultants Ltd. AOSERP Report 52. 209 pp.
- Davison, D.S., C.J. Fortems, and K.L. Grandia. 1976. Plume dispersion measurements from an oil sands extraction plant, March 1976. prep. for the Alberta Oil Sands Environmental Research Program by Intera Environmental Consultants Ltd. AOSERP Report 13. 195 pp.
- Desjardins, R.L. 1977. Description and evaluation of a sensible heat flux detector. *Boundary Layer Meteor.* 11: 147-154.
- Deuble, J.L. 1979. Atmospheric tracer gas technology, techniques and applications. Systems, Science and Software Technical Bulletin 79 - 5, La Jolla, CA. 11 pp.
- Dietz, R.N. and G.I. Senum. 1986. Capabilities, needs and applications of gaseous tracers. Atmospheric tracer technology and application. J.H. Heiken, ed. Noyes Publications p. 118 - 168.
- Eilts, M.D., A. Sundara-Rajan and R.J. Evans. 1987. Estimation of the average surface heat flux over an inhomogeneous terrain from the vertical velocity variance. *Boundary Layer Meteor.* 39: 87 - 91.

- Environment Canada, 1982. Canadian Climate Normals 1951-1980. Volume 5 Wind. Atmospheric Environment Service. Canadian Climate Centre. Downsview, Ontario. 283 pp.
- Fisher, N.I. 1987. Problems with the current definitions of the standard deviation of wind direction. *J. Clim. Appl. Meteor.* 26: 1522 - 1529.
- Fisher, N.I. 1983. Comment on "A method for estimating the standard deviation of wind directions". *J. Clim. Appl. Meteor.* 22: 1971.
- Frenzen, P. 1965. Determination of turbulence dissipation by Eulerian variance analysis. *Quart. J. Roy. Meteor. Soc.* 91: 28 - 34.
- Gill, C.G. 1975. Development and use of the Gill UVW anemometer. *Boundary Layer Meteor.* 8: 475 - 495.
- Harris, E.K. and R.A. McCormick. 1963. A simple procedure for estimating the standard deviation of wind fluctuations. *J. Appl. Meteor.* 2: 804 - 805.
- Hicks, B.B. 1972. Propeller anemometers as sensors of atmospheric turbulence. *Boundary Layer Meteor.* 3: 214 - 228.
- Hicks, B.B. 1970. The measurement of atmospheric fluxes near the surface: a generalized approach. *J. Appl. Meteor.* 9: 386-388.
- Hicks, B.B. and M.L. Wesley. 1978. An examination of some micrometeorological methods for measuring dry deposition. Argonne National Laboratory US EPA Report Number EPA-600/7-78-116. 29 pp.
- Holizki L.D. and Strom N. 1989. Field measurement of Atmospheric dispersion under low turbulence conditions for a well blowout or pipeline rupture. Paper presented to the 8th World Clean Air Congress. The Hague, Netherlands. 5. 49-54.
- Horst, T.W. 1973. Corrections for response errors in a three-component propeller anemometer. *J. Appl. Meteor.* 12(4): 716 - 725.
- Horst, T.W. 1972. A computer algorithm for correcting noncosine response in the Gill anemometer. Battelle Pacific Northwest Laboratory Annual for 1971. Report BNWL-1651 PT1, Richland, WA, USA. 183-188.
- Hyson, P. 1972. Cup anemometer response to fluctuating wind speeds. *J. Appl. Meteor.* 11: 843 - 848.
- James F. MacLaren Limited. 1978. Data source catalogue. Alberta Oil Sands Environmental Research Program. 170 pp.
- Johnson, W.B. 1985. Dispersion measurements using meteorological tracer techniques. Chapter 23. *Handbook of Applied Meteorology*, David D. Houghton, ed. John Wiley & Sons. New York, 679-696.
- Johnson, W.B. 1983. Meteorological tracer techniques for parameterizing atmospheric dispersion. *J. Clim. Appl. Meteor.* 22, 931 - 946.

- Jones, J.I.P. and F. Pasquill. 1959. An experimental system for directly recording statistics of the intensity of atmospheric turbulence. *Quart. J. Roy. Meteor. Soc.* 85: 225 - 236.
- Kaimal, J.C. 1986. Flux and profile measurements from towers in the boundary layer. *Probing the Atmospheric Boundary Layer*. 19 - 38. Amer. Meteor. Soc., Boston.
- Kaimal, J.C. 1981. Cup, propeller-vane and sonic anemometers in turbulence research. *Ann. Rev. Fluid Mech.* 13: 399 - 423.
- Kaimal, J.C. 1975. Sensors and techniques for direct measurement of turbulent fluxes and profiles in the atmospheric surface layer. *Atmos. Tech.* 7: 7 - 14.
- Kaimal, J.C. 1969. Measurement of momentum and heat flux variations in the surface boundary layer. *Radio Sci.* 4: 1147 - 1153.
- Kaimal, J.C. and D.A. Haugen. 1969. Some errors in the measurement of Reynolds stress. *J. Appl. Meteor.* 8: 460 - 462.
- Leahey, D.M. 1974. A study of airflow over irregular terrain. *Atmos. Environ.* 8: 783-791.
- Leahey, D.M. and M.C. Hansen. 1985a. Doppler acoustic radar observations of wind in the Elleslie region of Alberta. Prep. for Alberta Environment by Western Research, Division of Bow Valley Resources Services Ltd. RMD Report 85/31. 77 pp.
- Leahey, D.M. and M.C. Hansen. 1985b. Meteorological evidence for greater dispersion of elevated plumes. *J. Air Poll. Cont. Assoc.* 35: 828-830.
- Leahey, D.M. and M.J.E. Davies. 1978. Bivane studies of air flow over irregular terrain. *Proceedings of the 9th NATO International Technical Meeting on Air Pollution and its Application*. Toronto. pp. 89-100.
- Longley, R.W. 1956. The rate of ascent of pilot balloons and the variation of wind velocity as determined by them. *Suffield Technical Paper No. 96*. Defense Research Board. Dept. of National Defense, Canada.
- MacCready, P.B., Jr. 1966. Mean wind speed measurements in turbulence. *J. Appl. Meteor.* 5: 219 - 225.
- MacCready, P.B. Jr. 1965. Comparison of some balloon techniques. *J. Appl. Meteor.* 4: 504 - 508.
- MacCready, P.B., Jr. and H. R. Jex. 1964a. Response characteristics and meteorological utilization of propeller and vane wind sensors. *J. Appl. Meteor.* 3: 182 - 193.
- MacCready, P.B., Jr. and H.R. Jex. 1964b. Turbulent energy measurements by vanes. *Quart. J. Roy. Meteor. Soc.* 90: 198 - 203.
- Markee, E.H., Jr. 1963. On the relationships of range to standard deviation of wind fluctuations. *Monthly Weather Review*, 91(2): 83 - 87.

- Mathews, T. and R.B. Hicks. 1984. Doppler sodar measurement of wind climatology of Calgary. Prep. for Alberta Environment, Research Management Division by Department of Physics, The University of Calgary. RMD Report 84/28. 149 pp.
- Mathews, T. and R.B. Hicks. 1978. A study of the stability of the atmospheric boundary layer as revealed by an acoustic sounder and its relation to pollution levels in Calgary. Report submitted to Research Secretariat, Alberta Environment. 87 pp.
- McBean, G.A. 1972. Instrument requirements for eddy correlation measurements. *J. Appl. Meteor.* 11: 1078 - 1084.
- Misra, P.K. 1979. Determination of boundary layer - parameters using a vertical Gill anemometer. *Boundary Layer Meteor.* 17: 93 - 99.
- Middleton, W.E.K. and A.F. Spilhaus. 1953. *Meteorological Instruments*. University of Toronto Press.
- Mooney, M. L. and R. N. Swanson. 1976. Range statistics as estimators of intensity of turbulence at Diablo Canyon. *J. Appl. Meteor.* 15: 846 - 851.
- Mori, Y. 1987. Methods for estimating the mean and the standard deviation of wind direction. *J. Clim. Appl. Meteor.* 26: 1282 - 1284.
- Mori, Y. 1986. Evaluation of several "single-pass" estimators of the mean and the standard deviation of wind direction. *J. Clim. Appl. Meteor.* 25: 1387 - 1397.
- Moses, H. 1968. Meteorological instruments for use in the atomic energy industry. Chap. 6. *Meteorology and Atomic Energy*, D. H. Slade, editor, U.S. AEC. 257-300.
- Murray, W.A. 1983. Statistical analysis of wind and temperature profiles at Mildred Lake. prep. by Promet Environmental Group for Syncrude Canada Ltd. December 1983. 33 pp.
- Murray, W.A. 1977. Probabilities of emission limitation - analysis of meteorological data. Syncrude professional paper 1977-2. 12 pp.
- Murray, W. and J. Kurtz. 1976. A predictive study of the dispersion of emissions from the Syncrude Mildred Lake plant. Syncrude Canada, Environmental Research Monograph 1976-1. 128 pp.
- Murrow, H.N. and R.M. Henry. 1965. Self - induced balloon motions. *J. Appl. Meteor.* 4: 131 - 138.
- Nelson, E.W. 1984. A simple and accurate method for calculation of the standard deviation of the horizontal wind direction. *J. Air Poll. Cont. Assoc.* 34: 1139 - 1140.
- Netterville, D.D.J. 1981. User experience with a commercial acoustic Doppler remote wind sensor. *Proceedings of the International Symposium on Acoustic Remote Sensing of the Atmosphere and Oceans*, June 1981, VII - 11 to VII - 18. Printed by University of Calgary Printing Services Library of Congress Catalogue No. ISBN: 0-88953-031-9 1982.

- Netterville, D.D.J. and S.G. Djurfors. 1979. Controlling inherent uncertainties in double theodolite measurements. *J. Appl. Meteor.* 18: 1371 - 1375.
- NRC (Nuclear Regulatory Commission). 1972. Onsite meteorological programs. NRC Regulatory Guide 1.23, Washington, D.C.
- Panofsky, H.A. 1963. Determination of stress from wind and temperature measurements. *Quart. J. Roy. Meteor. Soc.* 89: 85 - 94.
- Pasquill, F. 1974. *Atmospheric Diffusion*. 2nd edition. Halsted Press. John Wiley & Sons, New York. 429 pp.
- Panofsky, H.A. and J.A. Dutton. 1984. *Atmospheric turbulence: models and methods for engineering applications*. Wiley Interscience. New York. 397 pp.
- Pendergast, M.M. and T. V. Crawford. 1974. Actual standard deviations of vertical and horizontal wind direction compared to estimates from other measurements. Preprint Symp. Atmos. Diff. Air Poll. Santa Barbara, Sept. 9 - 13, 1974. *Amer. Meteor. Soc.* 1-6.
- Rider, L.J. and M. Armendariz. 1966. A comparison of tower and pibal wind measurements. *J. Appl. Meteor.* 5: 43 - 48.
- Sackinger, P.A., D.D. Reible, and F.H. Shair. 1982. Uncertainties associated with the estimation of mass balances and gaussian parameters from atmospheric tracer studies. *J. Air Poll. Control Assoc.* 32: 720-724.
- Sakiyama, S.K. and R.P. Angle. 1988. Modelling - tracer study for risk assessment of a proposed sour gas facility. *Environ. Mon. and Assess.* 10-2: 133-46.
- Sakiyama, S. and J. Wong. 1987. Summary of data from the Ellerslie Meteorological Station 1978 - 1985. *Alberta Environment*. October 1987. 120 pp.
- Sakiyama, S.K. 1983. Numerical modelling and field verification for the Shell Moose Mountain Sour Gas project. *Alberta Environment*. Feb. 1983. 98 pp.
- Scoggins, J.R. 1965. Spherical balloon wind sensor behavior. *J. Appl. Meteor.* 4: 139 - 145.
- Schwiesow, R.L. 1986. A comparative overview of active remote-sensing techniques. *Probing the Atmospheric Boundary Layer*. 129 - 138. *Amer. Meteor. Soc.*, Boston.
- Teunissen, H. W., K. O. Vanek and S. E. Gould. 1985. Field intercomparisons of a propeller and sonic anemometer as sensors of atmospheric turbulence. Research Report NO. MSRB-85-5, Atmospheric Environment Service, Downsview, Ontario, Canada.
- Thyer, N. 1962. Double theodolite PIBAL evaluation by computer. *J. Appl. Meteor.* 1, 66-68.

- Turner, D.B. 1986. Comparison of three methods for calculating the standard deviation of the wind direction. *J. Clim. Appl. Meteor.* 25: 703-707.
- Turner, J.A. 1968. Standard deviation of wind direction estimated from direct observation of a sensitive wind vane. *J. Appl. Meteor.* 7: 714-715.
- Vanderborght, B. and J. Kretzchmar. 1984. A literature survey on tracer experiments for atmospheric dispersion modelling studies. *Atmos. Environ.* 18, 2395 - 2403.
- Verrall, K.A. and R.L. Williams. 1982. A method for estimating the standard deviation of wind directions. *J. Appl. Meteor.* 21: 1922 - 1925.
- Walker, E.R. 1964. Atmospheric turbulence characteristic measured at Suffield Experimental Station. *Beiträge zur Physik der Atmosphäre.* 37:38 - 52.
- Wagonner, P.E. 1965. *Agricultural Meteorology, Meteorological Monographs.* 6, pp. 28. Amer. Meteor. Soc., Boston, MA.
- Western Research and Development. 1976. Application of an air column trajectory pollution model to the City of Edmonton, Volume 2 (Appendices). Alberta Environment. 88 pp.
- Wieringa, J. 1980. Representativeness of wind observations at airports. *Bull. Amer. Meteor. Soc.* 61: 962 - 971.
- Woodruff, R.K. and C.S. Glantz. 1985. Survey of meteorological and tracer data for demonstrating a data archive. Battelle Pacific Northwest Labs. EPA/600/3-85/052 PB 85-235190. 53 pp.
- World Meteorological Organization. 1972. Guide to meteorological instrument and observing practices. Section 1.3.2: Site and exposure of instruments. WMO No. 8 TP 3 (revised ed.). I.7.
- Wyngaard, J.C. and S.F. Clifford. 1978. Estimating momentum, heat and moisture fluxes from structure parameters. *J. Atmos. Sci.* 35: 1204 - 1211.
- World Meteorological Organization. 1971. Observational requirements of lower tropospheric soundings. *Boundary Layer Meteor.* 2: 128 - 129.
- Yamartino, R.J. 1984. A comparison of several "single-pass" estimators of the standard deviation of wind direction. *J. Clim. Appl. Meteor.* 23: 1362 - 1366.

APPENDIX B

DATA PRESENTATION TECHNIQUES

B.1 WIND DIRECTION

In summaries of wind direction data, using the 16- or 8-point compass, it is often necessary to convert from azimuth degrees. This is awkward for data collected in tens of degrees because neither 16 nor 8 are factors of 36 (the number of possible values in the tens of degrees format). Such conversions necessarily involve arbitrary conventions as shown in Table B.1). In each conversion the sectors corresponding to the cardinal points are larger than the sectors for the non-cardinal points. Moreover, in the 16-point conversion the centrelines of alternate non-cardinal sectors are shifted away from the true direction (Lea and Helvey 1971). As a result, the directional frequency distributions of converted data are biased. The frequencies of cardinal directions are systematically enlarged from their true values while the frequencies of non-cardinal directions are systematically reduced. Approximate corrections that can be applied as an operational remedy are given in Table B.2 (Lea and Helvey 1971).

Although the bar wind rose is the most common, polygonal or continuous wind roses are sometimes used. These are constructed by joining the ends of the radii with straight line segments or by drawing a smooth curve through all the end points as in Figure B.1 a, b. Wind roses are really two-dimensional representations of a three-dimensional function. No mapping that involves a change in the number of dimensions can preserve all the attributes of a closed figure, namely, distance, area and shape. In bar wind roses distance is preserved at the expense of the other two qualities, by making the radii proportional to the frequencies (Court 1963).

A basic principle of graphics is that each element of the diagram should cover the entire range to which it applies. In an 8-point wind rose, each frequency applies to winds coming from a sector of 45° ; therefore each frequency should be represented by a figure covering the entire 45° and not just a small central portion. The conventional bar rose may give an erroneous picture of the actual wind distribution. A wedge rose (Figure B.1 c) would be superior.

For most purposes an area-preserving transformation is preferable to the common distance-preserving one (Court 1963). It is the only type of transformation that yields a frequency diagram suitable for direct comparison with a theoretical distribution.

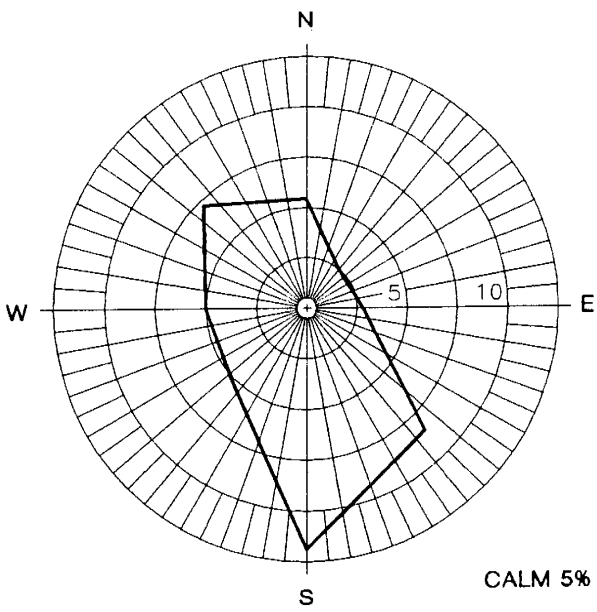
Table B.1 Usual method of converting from wind direction in tens of degrees to 16-point and 8-point compass.

Compass Point (16)	Direction in Tens of Degrees	Compass Point (8)	Direction in tens of Degrees
N	35, 36, 01	N	34, 35, 36, 01, 02
NNE	02, 03		
NE	04, 05	NE	03, 04, 05, 06
ENE	06, 07		
E	08, 09, 10	E	07, 08, 09, 10, 11
ESE	11, 12		
SE	13, 14	SE	12, 13, 14, 15
SSE	15, 16		
S	17, 18, 19	S	16, 17, 18, 19, 20
SSW	20, 21		
SW	20, 23	SW	21, 22, 23, 24,
WSW	24, 25		
W	26, 27, 28	W	25, 26, 27, 28, 29
WNW	29, 30		
NW	31, 32	NW	30, 31, 32, 33
NNW	33, 34		

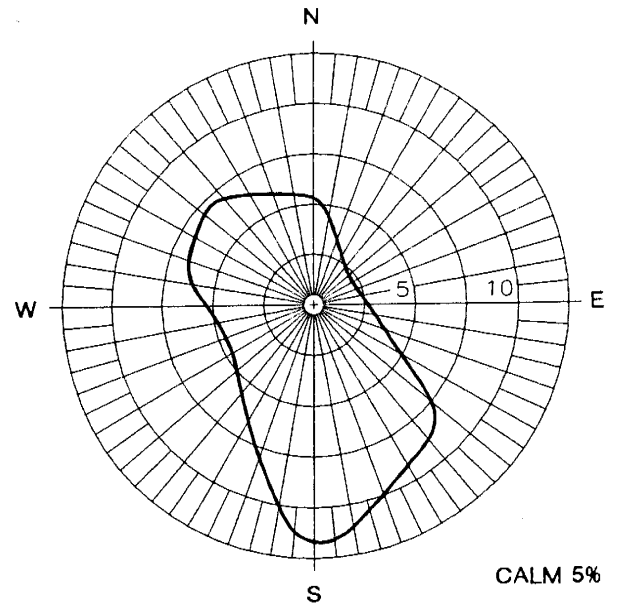
Table B.2 Debiasing procedure for direction frequencies from mixed compass formats (Lea and Helvey 1971).

Primed directions indicate adjusted frequencies; unprimed directions indicate apparent (unadjusted) frequencies.

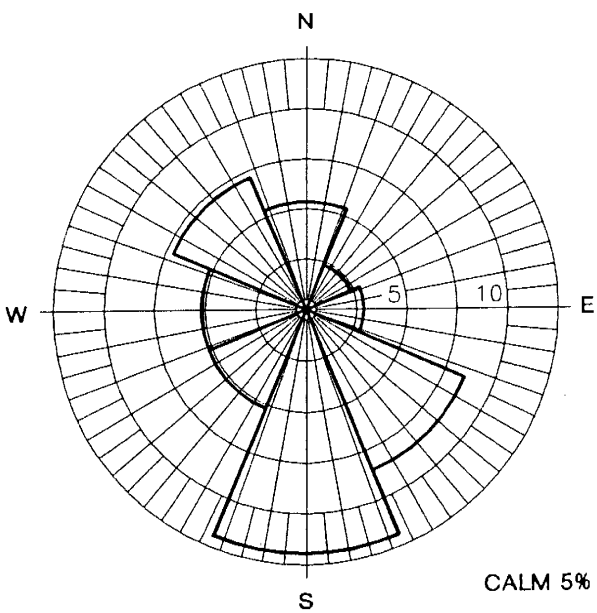
16-point directions	8-point directions
$N' = 0.75 N$	$N' = 0.9 N$
$NNE' = 0.9375 NNE + 0.125 N$	$NE' = NE + 0.05 N + 0.05 E$
$E' = NE + 0.0625 NNE + 0.0625 ENE$	$E' = 0.9 E$
$ENE' = 0.9375 ENE + 0.125 E$	$SE' = SE + 0.05 S + 0.05 E$
$E' = 0.75 E$	$S' = 0.9 S$
$ESE' = 0.9375 ESE + 0.125 E$	$SW' = SW + 0.05 S + 0.05 W$
$SE' = SE + 0.0625 ESE + 0.0625 SSE$	$W' = 0.9 W$
$SSE' = 0.9372 SSE + 0.125 S$	$NW' = NW + 0.05 N + 0.05 W$
$S' = 0.75 S$	
$SSW' = 0.9375 SSW + 0.125 S$	
$SW' = SW + 0.0625 SSW + 0.0625 WSW$	
$WSW' = 0.9375 WSW + 0.125 W$	
$W' = 0.75 W$	
$WNW' = 0.9375 WNW + 0.125 W$	
$NW' = NW + 0.0625 WNW + 0.0625 NNW$	
$NNW' = 0.9375 NNW + 0.125 N$	



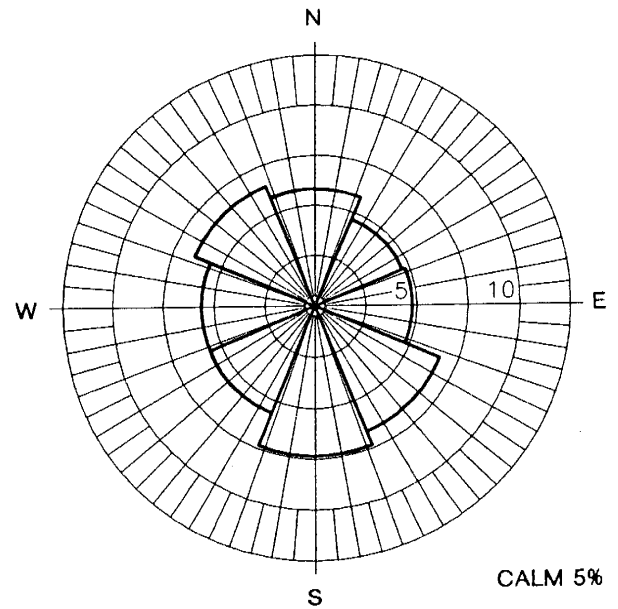
(a) polygonal



(b) continuous



(c) wedge



(d) sector area

Figure B.1 Different types of wind roses for Red Deer.

Such sector roses (Figure B.1 d) have radius lengths proportional to the square root of the wind frequency. When a central circle is desired to avoid drafting problems or to show calms, the radius length is proportional to $c^2 + f^2$ where c is the radius of the central circle and f is the frequency to be represented in the sector. Smooth curves can then be fitted empirically or analytically to provide a complete description of the wind direction distribution.

B.2 WIND SPEED

Wind speeds are most often tabulated as frequency of occurrence within various speed ranges that correspond to the Beaufort scale (Table B.3). The information for a number of Alberta aviation weather stations is given in Hawrelak et al. (1976).

If the Weibull distribution is used for the cumulative probability of wind speeds, several methods are available to estimate the Weibull parameters depending upon which wind statistics are available and the level of sophistication desired (Justus et al. 1978). Weibull probability graph paper (Figure B.2) is designed specifically for plotting wind speed distributions and estimating the Weibull distribution parameters to within a few percent (Takle and Brown 1978).

B.3 WIND VECTOR DATA (SPEED AND DIRECTION)

The bar wind rose used to display wind direction data can be easily modified to incorporate information about wind speeds. The bar representing the frequency in any given direction is simply divided into a number of segments with each segment proportional in length to the frequency of wind speeds in some arbitrary range. The segments are usually drawn with different widths or shadings for better visual separation (Figure B.3).

Such combined wind roses have a number of shortcomings (Patterson and Benjamin 1975, Graedel 1977, Crutcher 1957):

1. In order to express the extreme range of speeds, data must be compressed so that in some sectors few ranges will be displayed.
2. Quantitative information cannot be read without recourse to a length scale.

Table B.3 Beaufort wind speed equivalents (after WMO 1972).

Beaufort number	Wind Speed at 10 m (km/h)	Description	Observed Effects
0	<1	Calm	Calm; smoke rises vertically.
1	1-5	Light air	Smoke drifts slowly.
2	6-11	Light breeze	Wind felt on face; leaves rustle.
3	12-19	Gentle breeze	Leaves and small twigs in constant motion; wind extends light flag.
4	20-28	Moderate breeze	Raises dust and loose paper; small branches are moved.
5	29-38	Fresh breeze	Small trees in leaf begin to sway, crested wavelets form on inland waters.
6	39-49	Strong breeze	Large branches in motion; whistling heard in telegraph wires; umbrellas used with difficulty.
7	50-61	Near gale	Whole trees in motion; inconvenience felt when walking against the wind.
8	62-74	Gale	Breaks twigs off trees; generally impedes progress.
9	75-88	Strong gale	Slight structural damage occurs (chimney pots and slates removed).
10	89-102	Storm	Seldom experienced inland; trees uprooted; considerable structural damage occurs.
11	103-117	Violent storm	Very rarely experienced; accompanied by widespread damage.
12	>118	Hurricane	

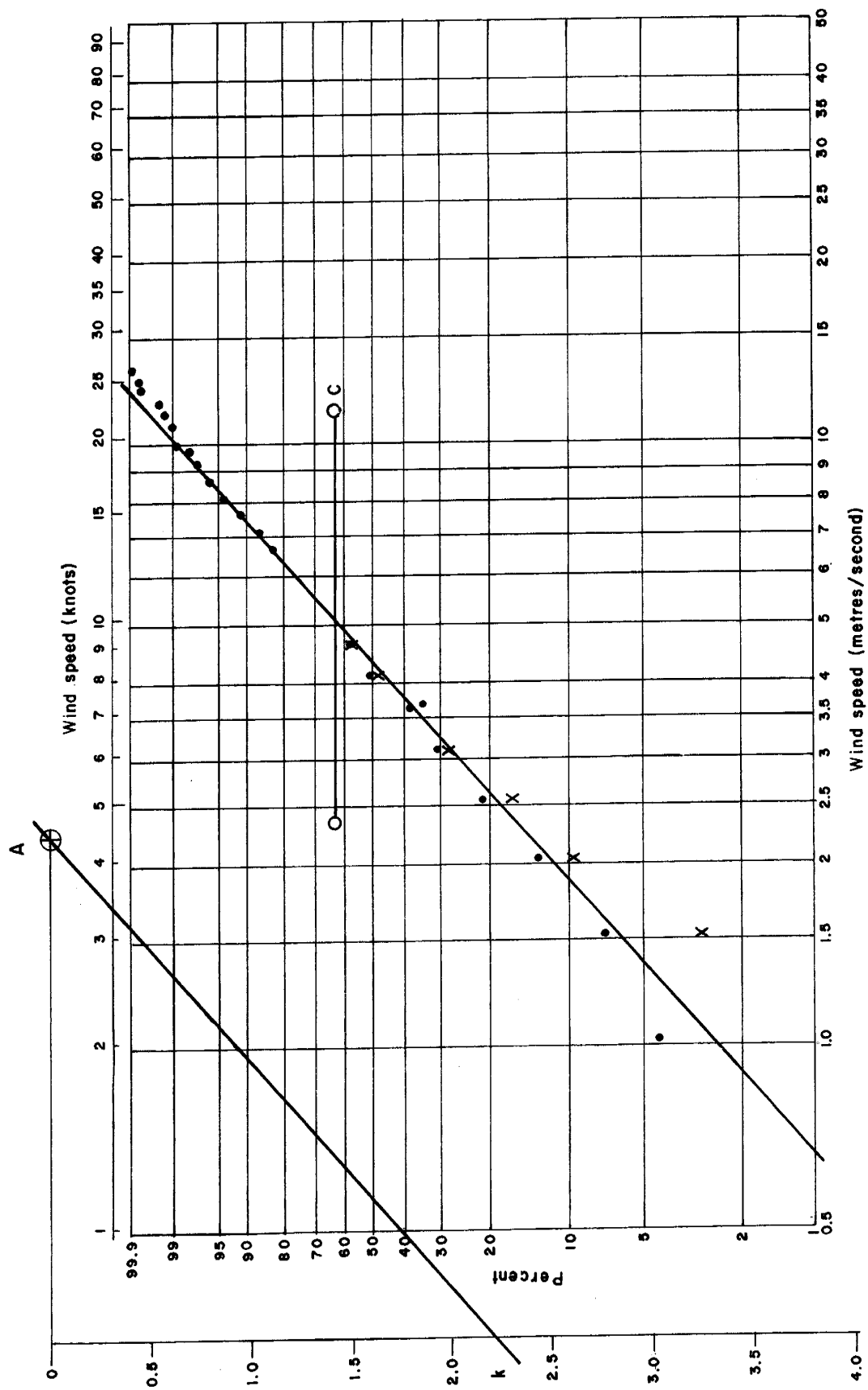


Figure B.2 Weibull probability paper (Takle and Brown 1978). The value of the scale factor C is given by the crossing of the 63.2 % level. The value of the shape factor k is determined by drawing a parallel line through point A to the point of intersection with the scale on the left.

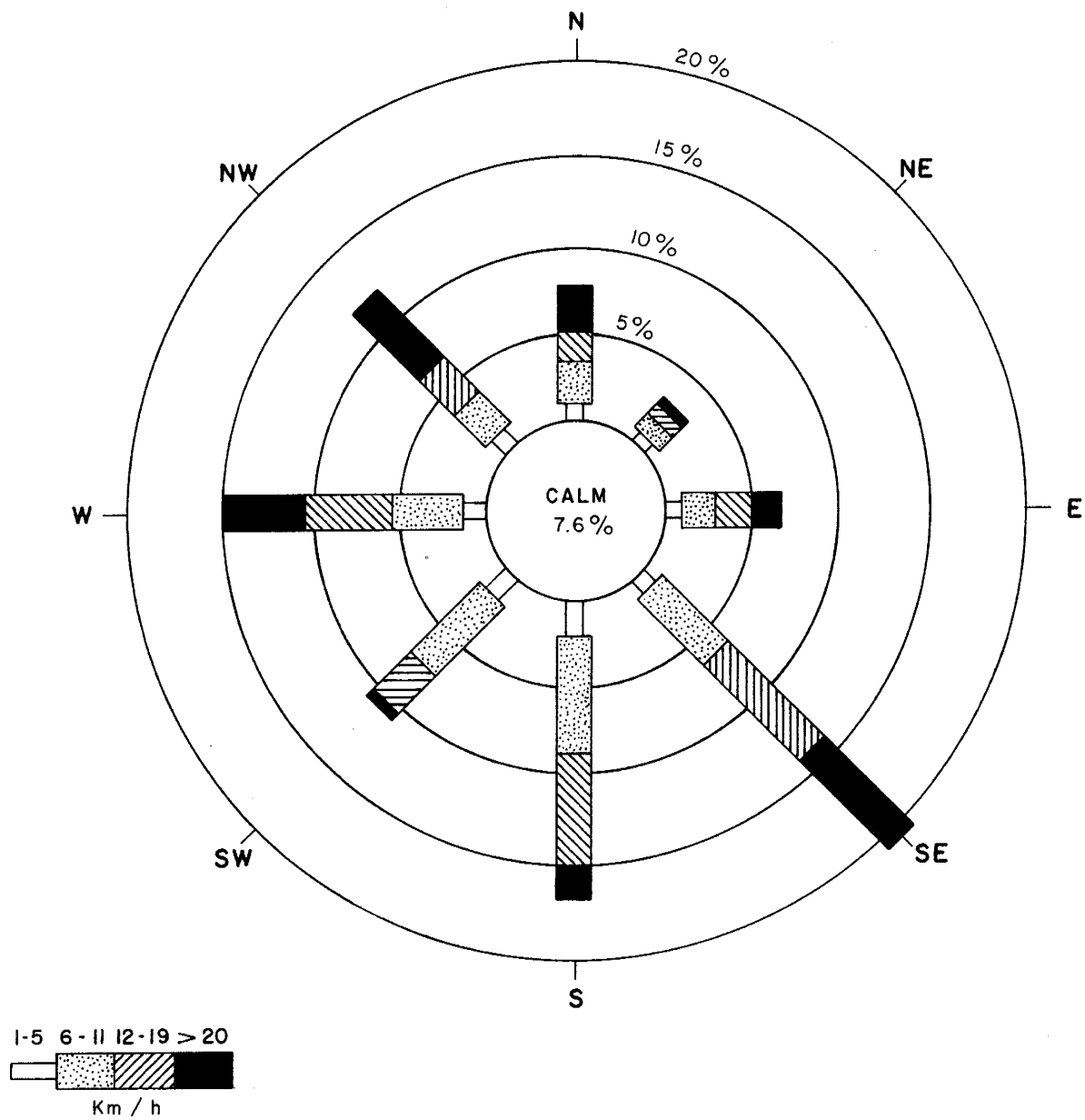


Figure B.3 Wind rose with speed segments, Edmonton International, winter 1961-1967.

3. The eye is a poor judge of radial distances and has difficulty comparing the distributions of wind speeds.
4. Interpolation between principal directions is difficult because of the divergence of the bars.
5. The important features of time sequences are difficult to discern.
6. The number of compass points and speed classes may be such that the user cannot readily interpret the diagram for this particular application.
7. Polar plots are tedious to prepare by hand and difficult to execute on many computer graphics systems.

A speed-direction frequency diagram allows the identification of any major wind systems and provides enough detailed information so that the controlling physical mechanisms may be postulated (Patterson and Benjamin 1975). To construct the diagram the wind frequencies are tabulated on squared paper. Each square or distribution cell represents a certain wind direction sector and speed class combination. Smooth isopleths of frequency are then drawn. (Figure B.4).

The "boxplot", a statistical technique for the simultaneous presentation of several distributions of data, can also be used (Graedel 1977). Selected characteristics of the speed – usually the highest value, upper quartile, median, lower quartile and lowest value – are computed and plotted as a function of direction (Figure B.5). The discontinuity at North is overcome by duplicate plotting of the first few sectors. The frequency of wind flow in each sector is indicated by a figure in the box. Such plots give the quantitative details of the wind speed distribution in each sector and allow immediate detection of small differences. To avoid small numbers in the samples for some sectors, the plot can be made with sector centres adjusted so that sample sizes are equal; then no numerical frequency information has to be added to the display.

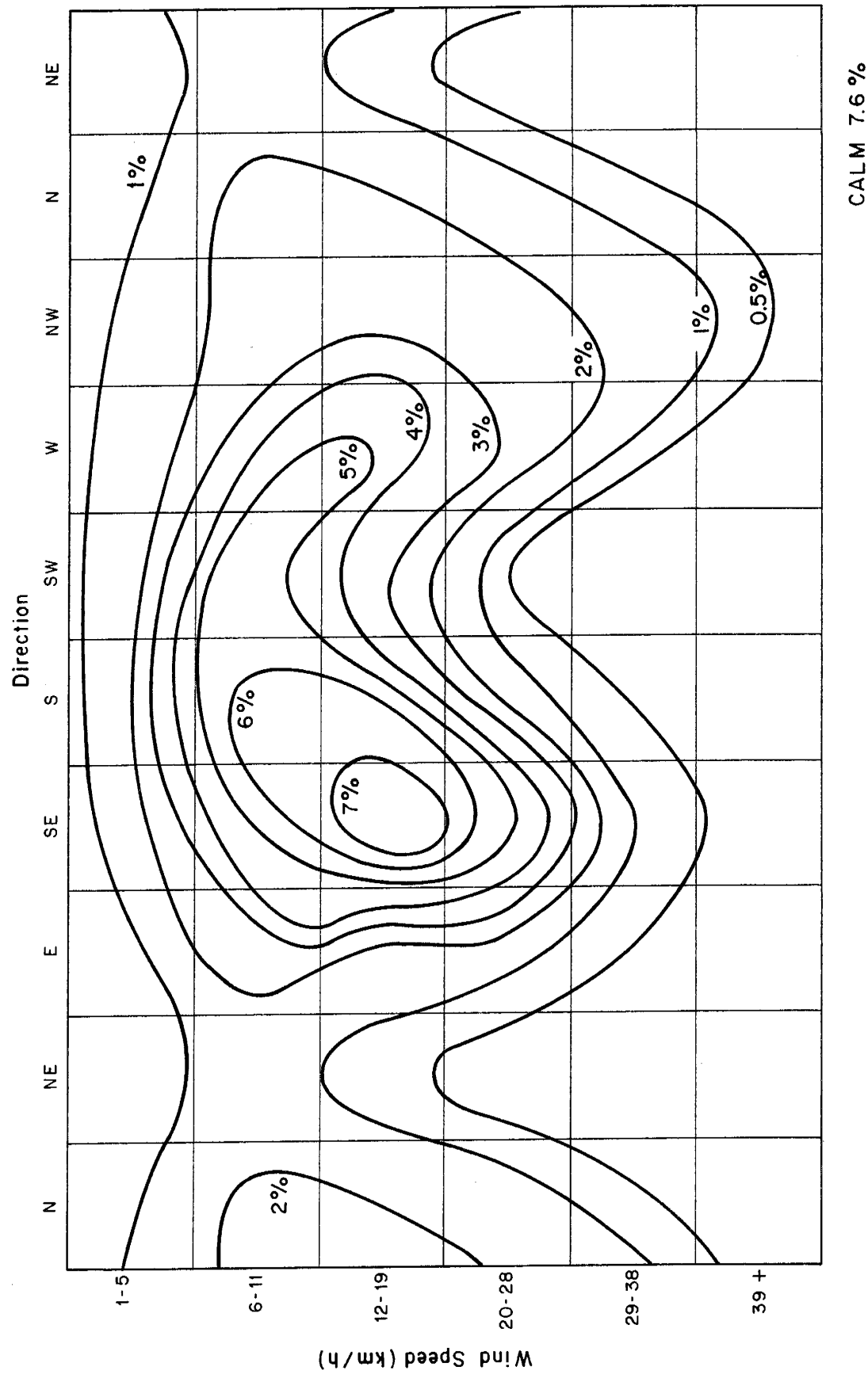


Figure B.4 Speed-direction frequency diagram, Edmonton International, winter 1961-1967.

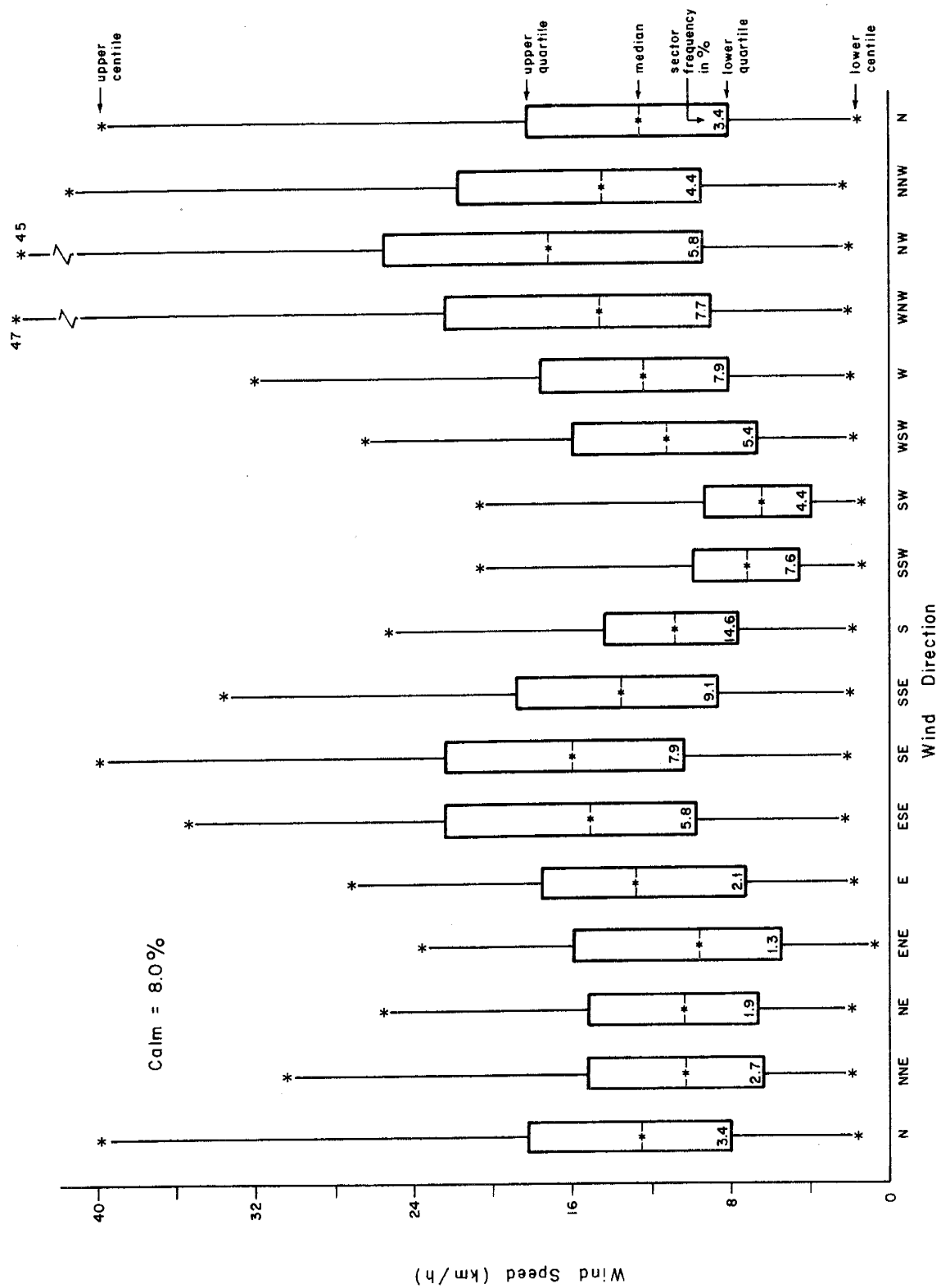


Figure B.5 Boxplot for wind at Edmonton International, winter 1971-1975.

The standard vector-deviation wind rose represents the wind distribution by two parameters – the vector mean and the vector standard deviation (Crutcher 1957). For these two parameters any conventional wind rose can be reconstructed. The standard vector deviation is computed by means of the equation

$$\sigma_V^2 = \sum \frac{V^2}{n} - V_{avg}^2$$

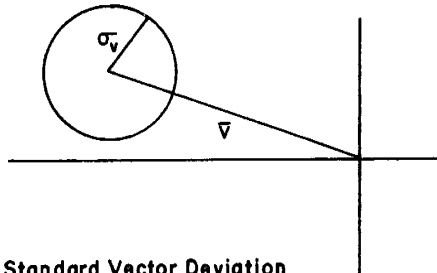
where $\sum V^2$ is the sum of squares of individual speeds and V_{avg}^2 is the square of the vector mean wind speed. The first step in producing the diagram consists of plotting the vector mean wind with the origin of the coordinate system as the head of the vector. With the tail of the vector as the centre, a circle is drawn with radius equal to the standard vector deviation. The area encompassed by the circle contains the origin of 63% of the distributed winds. Circles corresponding to other percentiles can also be drawn. Their radius is calculated as the product of σ_V and the value obtained from Table B.4. For example, a radius of $0.83 \sigma_V$ describes a circle which encloses 50% of the winds. The manner of extracting information from such a diagram is illustrated in Figure B.6.

If the observed winds come from distinctly different regimes, the distribution may not be circular. An elliptical distribution characterized by a major and minor axis, standard vector deviation and a correlation coefficient may be more appropriate. In situations where the wind pattern is quite distorted, neither diagram will adequately describe the data. This is often the case for surface winds. The standard vector deviation rose is probably best applied to upper level winds.

The polar isopleth diagram (Bower and Sullivan 1981) is prepared by plotting the frequency of occurrence of winds in cells defined by ranges of radial and angular components. Lines are then drawn connecting cells of equal frequency (Figure B.7). Chemical concentration could also be plotted to show the relationship between wind and air quality. Wind fabric diagrams (Davis and Ekern 1977) use a stereographic projection that preserves area.

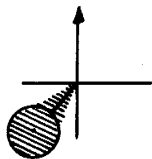
Table B.4 Normalized radius of the percentile circle in standard vector deviation rose (Crutcher 1957).

Area contained in circle (P)	Radius/standard vector deviation (R/ σ_v)
0.05	0.22
0.10	0.32
0.15	0.40
0.20	0.48
0.25	0.54
0.30	0.59
0.35	0.66
0.40	0.72
0.45	0.78
0.50	0.83
0.55	0.89
0.60	0.96
0.63	1.00
0.70	1.10
0.75	1.18
0.80	1.27
0.85	1.38
0.90	1.52
0.95	1.73
0.99	2.15

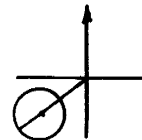


(a) Vector Mean Wind and Standard Vector Deviation

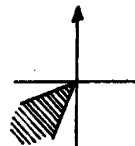
The magnitude of the vector mean wind speed is shown by the distance from the origin to the centre of the circle. The radius of the circle is equal to one standard vector deviation and encompasses 63% of all wind vector tails.



(b) Minimum Wind Speeds The magnitude of the minimum wind speed for the central 63% distribution is shown by the smallest distance from the compass centre to the 63% distribution circle.



(c) Maximum Wind Speeds The magnitude of the maximum wind speed for the central 63% distribution is shown by the greatest distance from the compass centre to the 63% distribution circle.



(d) More than 63% of the time winds will be observed to have a direction from that portion of the compass indicated by the shaded sector area.

Figure B.6 Schematic illustration as to how to extract information from graphical presentation of circular standard vector deviation wind roses. This procedure is applicable to elliptical distributions.

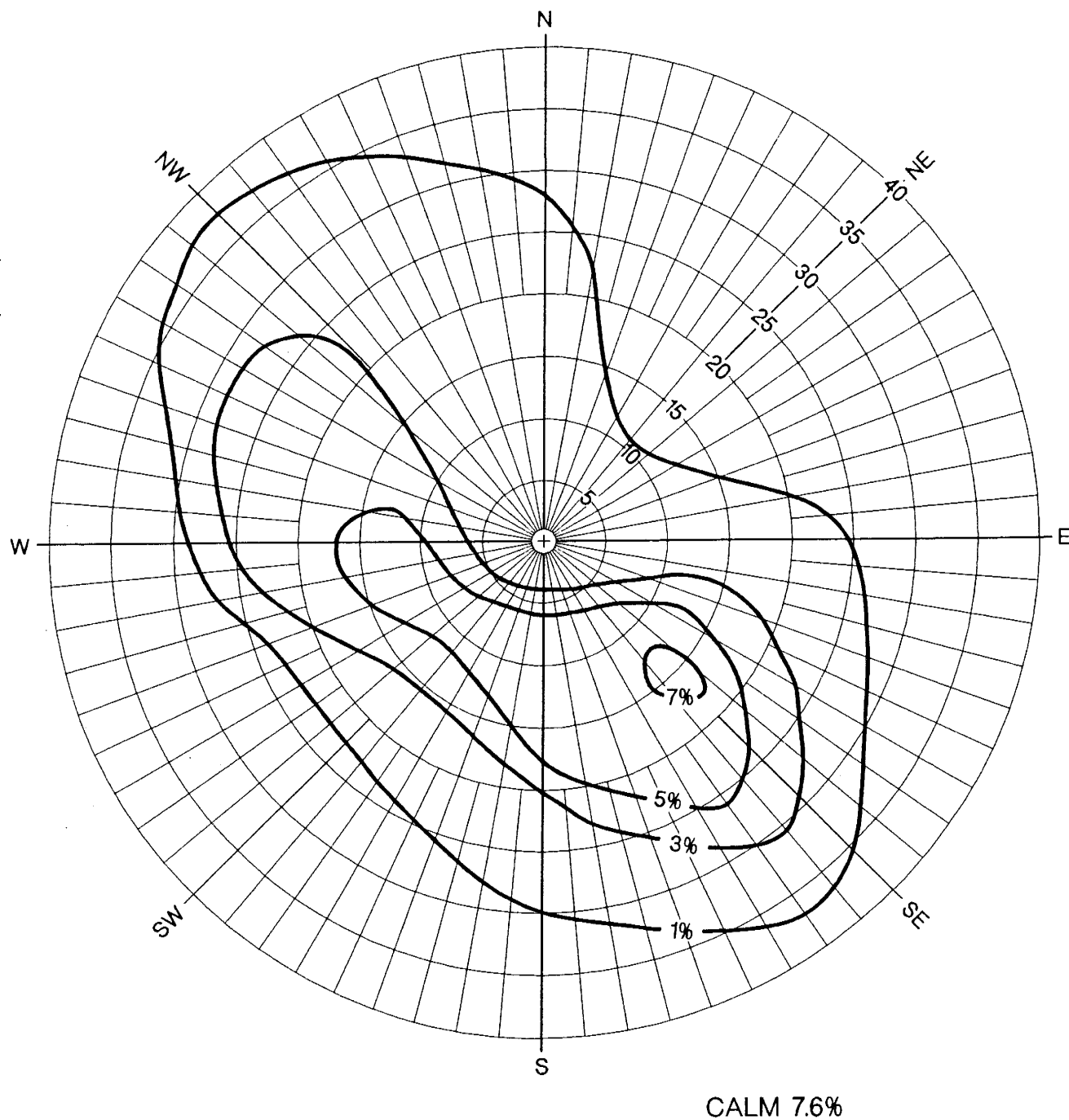


Figure B.7 Polar isopleth diagram, Edmonton International airport, Winter 1961-1967.

B.4 REFERENCES

- Bower, J.S. and E.J. Sullivan. 1981. Polar isopleth diagrams: a new way of presenting wind and pollution data. *Atmos. Environ.* 15: 537-540.
- Court, A. 1963. Wind roses. *Weather* 18: 106 - 110.
- Crutcher, H. L. 1957. On the standard vector-deviation wind rose. *J. Meteor.* 14: 28 - 33.
- Davis, B.L. and M.W. Ekern. 1977. Wind fabric diagrams and their application to wind energy analysis. *J. Appl. Meteor.* 16: 522-531.
- Graedel, T.E. 1977. The wind boxplot: An improved wind rose. *J. Appl. Meteor.* 16: 448 - 450.
- Hawrelak, J., T. Rachuk and J. Barlishen. 1976. *Wind Power in Alberta*. Alberta Research Council, Edmonton. 110 pp.
- Justus, C. G., W. R. Hargraves, A. Mikhail and D. Graber. 1978. Methods for estimating wind speed frequency distributions. *J. Appl. Meteor.* 17: 350 - 353.
- Lea, D. A. and R. A. Helvey. 1971. A directional bias in wind roses due to mixed compass formats. *J. Appl. Meteor.* 10: 1037 - 1039.
- Paterson, M.P. and S.F. Benjamin. 1975. Better than a wind rose. *Atmos. Environ.* 9: 537 - 542.
- Takle, E.S. and J.M. Brown. 1978. Note on the use of Weibull statistics to characterize wind speed data. *J. Appl. Meteor.* 17: 556 - 559.
- WMO (World Meteorological Organization). 1971. *Guide to meteorological instrument and observing practices*. Section 1.3.2.: Site and exposure of instruments. WMO No. 8 TP 3 (revised ed.). I.7.

PLUME DISPERSION IN ALBERTA

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This book arose from the needs of environmental specialists for both an explanation of basic plume concepts and a digest of recent advances in knowledge, particularly from studies in local terrain and climate.

Topics include: elements of dispersion meteorology; characteristics of the atmospheric boundary layer; principles governing plume rise; plume dispersion over gentle terrain; flow and diffusion over rugged terrain; atmospheric measurement methods; and data presentation techniques. Theoretical and experimental investigations in Alberta are reviewed and summarized. The more than 400 pages of text use 98 figures and 58 tables to illustrate concepts and report Alberta work. The unique qualities of the book are its placement of atmospheric science within the Alberta context, and its demonstration of the contribution of Alberta studies to general understanding of plume dispersion.

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