



LIBRARY-AIR RESOURCES BOARD

CONTRACT NO. A932-051

FINAL REPORT

OCTOBER 1991

**Analysis of the
1987 Southern California
Air Quality Study (SCAQS)
Sulfur Hexafluoride Atmospheric
Tracer Data**

A series of three parallel diagonal stripes, consisting of a light gray band, a dark gray band, and a white band, running from the bottom left towards the top right of the page.

**State of California
AIR RESOURCES BOARD
Research Division**

**Analysis of the 1987
Southern California Air Quality Study (SCAQS)
Sulfur Hexafluoride Atmospheric Tracer Data**

**Final Report
Contract No. A932-051**

Prepared for:

Research Division
California Air Resources Board
1800 15th Street
Sacramento, CA 95814

Submitted by:

California Institute of Technology
Chemical Engineering Department
Pasadena, California

Prepared by:

Frederick H. Shair
Principal Investigator

LIBRARY
CALIFORNIA AIR RESOURCES BOARD
P.O. BOX 2815
SACRAMENTO, CA 95812

OCTOBER 1991

LIBRARY
CALIFORNIA AIR RESOURCES BOARD
P.O. BOX 2815
SACRAMENTO, CA 95812

Table of Contents

	Page
Abstract	3
Introduction	4
Results	7
Summary and Conclusions	8
Appendix -	
PhD. Thesis: Aspects of Atmospheric	Attached
Transport and Dispersion Within an Air Basin	
by Robin Scott Horrell	

Abstract

Eight sets of atmospheric tracer experiment data obtained during the 1987 Southern California Air Quality Study (SCAQS) are analyzed by mass balance and qualitative agreement with surface winds gathered during the same period. The sulfur hexafluoride tracer releases done near downtown Los Angeles reveal aspects of the complexity of the atmospheric emission transport from this location. The total material release of sulfur hexafluoride can be accounted for by mass balancing. The mass balances are used to examine the residence time of the tracer in the basin. The residence time for the sulfur hexafluoride tracer released from Vernon CA is 10 hours in the summer, and exceeds 24 hours for the fall. The primary exit route of the sulfur hexafluoride tracer during the summer releases was the San Fernando Valley to the northwest of downtown.

The surface-level diagnostic wind model (DWM) was made operational on a PC. A mechanism for accepting the output from the DWM into WIND, the Caltech PC-based wind field interpretation tool was devised. A detailed manual describing the operation of WIND, and the PC version of WIND2D was developed, and provided to the CARB under separate cover.

Introduction

The Southern California Air Quality Study (SCAQS) of 1987 was an extensive multi-party effort to gather the necessary data to develop the tools used for air quality management in the Los Angeles Basin. The development of tools is an iterative process involving model formulation, experiments, and computer simulation, each building upon the other, and becoming increasingly refined. As part of the experimental portion of the SCAQS, atmospheric tracer experiments were conducted. Well executed gaseous tracer experiments are a cornerstone for the testing and development of atmospheric transport and dispersion models. Tracer studies also serve as concrete evidence of the transport and dispersion which occurred on specific days, and can help investigators to develop conceptual models regarding transport and dispersion in domains of complex air flow. During the SCAQS two types of tracer experiments were conducted. One used sulfur hexafluoride and the other four separate perfluorocarbons. Each test augmented the other by using different tracer gases, and allowing different locations in the basin to be studied simultaneously.

The sulfur hexafluoride experiment was designed to provide a controlled method for assessing the transport and dispersion of a single point source from selected sites and for specific times during intensive SCAQS days. This is consistent with the third goal and objective described by Blumenthal et al. (1989):

3. to enhance our understanding of the relationship between emissions and the spatial and temporal distributions of pollutants so that air quality simulation models and, ultimately, air quality management strategies can be improved.

The tracer experiments were designed to examine both the short and long distance impact of surface emissions within the basin. The sulfur hexafluoride tracer experiments had two specific objectives: (1) to label emissions into the early morning and evening wind direction transition periods characteristic of coastal environments, and (2) to identify the source-receptor relationships of surface level emissions originating from near downtown Los Angeles.

Preliminary analysis of the tracer data was done to ensure that the tracer was in fact observed by the respective groups which conducted the experiments. The original analysis of the tracer data showed it could be used to examine source-receptor relationships, and to characterize transport and dispersion patterns. The sulfur hexafluoride tracer data revealed a number of unexpected findings regarding source-receptor relationships.

The transport of the tracer is generally described as a trajectory showing the location of the tracer as a function of time and spatial position. In order to visualize the tracer's trajectory and the wind fields responsible for transporting the tracer a PC-based visualization and processing tool, WIND, has been developed at Caltech. One goal of this project was to compare observed tracer transport with the transport which would be predicted based on meteorological data. A second aspect of this project was to bridge the gap between simulation and visualization of wind fields. A part of this contract was to interface to the diagnostic wind model to WIND, and to describe the general interface which exists between WIND and all wind field models. Because early assessments about the ease of porting more complex airshed models established that limitations to this existed, research plans were adjusted accordingly. This work focused primarily on the interfacing of the output of airshed models to the visualization, and data processing capability of WIND. As part of the data processing aspect, mass balances on the tracer were done to ensure that the major portion of the tracer was accounted for by the measurements.

The research performed under this grant also pertains to comparing the measured tracer transport patterns to wind fields gathered during the release days. The transport patterns themselves were interpreted for the sulfur hexafluoride tracer experiments because of the especially unexpected results which were obtained. These are described in detail in Section 2 of Appendix A.

An important objective of this work was to assess whether existing surface wind models are capable of describing the observed tracer trajectories. Where possible

this issue was addressed.

Results

1. The diagnostic wind model for surface winds was ported to the PC.
2. It was impractical to port the 3-dimensional diagnostic wind model to the PC for 2 reasons: (1) Fortran routines, that were system specific, had been used within the model and were impractical to redevelop for the PC, (2) the PC's memory capacity was exceeded by the larger 3-D model. (See Appendix B.1 for more details). At the current rate in improvements of PC and workstation technology reevaluation of these limits in two years would be appropriate.
3. A general interface to the PC-based visualization tool WIND was described, and a specific interface of the diagnostic wind model to WIND was created, and tested.
4. The WIND tool was used to show the diagnostic wind model (WIND2D) is inadequate at describing the summertime sulfur hexafluoride tracer releases from Vernon. The DWM was insensitive to the location of the diurnal divergence zone located between downtown Los Angeles and Glendale. (Refer to Section 2. of Appendix A which describes this divergence zone in detail). The model was tested using a standard input data from 74 meteorological stations within the modeling grid. This data was supplied by the ARB for the purpose of these runs.
5. The WIND package accurately traces the trajectory of summertime sulfur hexafluoride tracer releases from the Vernon, CA site. This substantiated WIND's sensitivity to divergence zones.
6. The algorithm in the WIND package consistently identifies the divergence zone using the raw surface winds. WIND was used to characterize the summertime Los Angeles - Glendale divergence zone to establish the statistical occurrence of the zone, and the extent and nature of the zone. (Refer to Section 2.3 of Appendix D). The zone divides the downtown Los Angeles emissions between two distinct exit routes from the basin.

7. The 1 sample per mile resolution provided by automobile traverses during the sulfur hexafluoride sampling period provided excellent spatial resolution of the tracer cloud.
8. The fixed sampling locations used for the SF_6 releases were located at an average of 7.2 ± 3 miles apart (or $11.5 \text{ km} \pm$) and provide only a coarse resolution for comparison with air shed models.
9. Sulfur hexafluoride tracer mass balances establish approximately 100 percent tracer accountability for summertime studies and from 50 to 80 percent accountability for fall studies. (Refer to Section 2.1 of Appendix D).

Summary and Conclusions

The following is a short list of general conclusions which were derived from this project. Following this list is an exhaustive list of specific conclusions which are drawn within Appendix A.

1. Personal computer based wind flow models are currently limited to those models which perform only limited data processing of the input meteorological data. This includes the surface based diagnostic wind model and WIND.
2. The PC is a perfect tool for visualizing wind fields because it has advanced graphics capabilities, is economical, and has wide spread use.
3. For diagnostic interpretation of wind fields within the Los Angeles Basin, where wind monitoring stations are relatively dense, the calculation algorithm used by WIND provides more realistic wind fields than the diagnostic wind model which was tested.
4. The diagnostic wind model (DWM) provides useful information about the wind fields on the perimeter of the basin where wind sampling is sparse.
The WIND model should be used in future field studies, especially those where (1) wind data is available from a telemetry system, and (2) any tracer is being released. This will allow investigators to focus their sample gathering efforts in areas which are expected to be impacted by the tracer.
5. Using WIND as a real time guide, future sulfur hexafluoride tracer studies might be conducted more extensively with automobile traverses and therefore yield higher spatial resolution of the tracer clouds.

The following set of conclusions was extracted from R. S. Horrell's thesis in the order they appeared. Section 2 of the thesis was a direct result of the work funded by this grant. Sections 1 and 3 are on related subjects that were studied in the area of air quality and atmospheric studies: (1) thermally driven slope wind and (3) ozone forecasting using a neural network. Please refer to the thesis as needed.

R. S. Horrell's Thesis Conclusions by Section:

Conclusions Sections 1.1 - 1.6

Probably the most striking feature of buoyantly driven slope flows brought out by this analysis is the functional relationship between the boundary layer thickness and the slope length (1.6.2a and 1.6.2b). The two regimes are controlled by the mechanisms by which energy flows across the buoyant layer. At the leading edge of the slope, advective and diffusive transport balance, while further along the slope diffusive transport is balanced by the heating (or cooling) effect of entraining ambient air into the momentum boundary layer. The demarcation point between the two regimes is set by the value of the parameter λ . For $\lambda \leq 1$ the former case above holds, while for $\lambda \geq 1$ the latter case applies. Of course, in the overlap region around $\lambda = 1$ all three mechanisms for energy transport are in balance. It is very important when assessing the potential for slope flow to first calculate λ . The steeper and shorter the slope or the smaller the ground to air temperature gradient, the more likely λ will exceed 1. Based on λ 's value, one can calculate the remaining parameters of interest.

The growth of the buoyantly driven boundary layer at its leading edge as a function of to the quarter power of the length along the slope ($\Delta \sim L^{\frac{1}{4}}$) is one aspect that has important implications with regard to pollution transport. The reason for this lies in the connection between boundary layer thickness and the entrainment of air from above the layer. A boundary layer which grows as $\chi^{\frac{1}{4}}$, where χ is used instead of L since one can choose to scale the slope length any

where along the slope, entrains a large fraction of air at first and less and less with increasing distance along χ . This means that a buoyant boundary layer acts like a pump, preferentially extracting air at its leading edge and depositing it at the end of the slope.

This has two major effects on the transport of pollutants, depending on their origin relative to the buoyant flow. First, if elevated pollution levels exist at the front of the buoyant boundary layer, the polluted air is drawn in and transported within the layer until the end of the slope is reached. Dilution of the pollutant by air from above the layer farther along the slope is limited because the slow growth of the layer imposed by the $\chi^{\frac{1}{4}}$ law provides minimal entrainment into the layer. If the slope is long enough such that λ exceeds 1 ($\Delta \sim L^0$), then the boundary layer thickness becomes constant, ie., no entrainment into the layer from above, and dilution is limited completely to turbulent transport across the boundary layer. Second, if the slope along which the boundary layer develops is a source of pollutant, for instance a landfill, then by the same arguments presented above, the pollutant is trapped and transported within the boundary layer. In this way the emissions from a ground source can be moved at groundlevel from one location with limited dilution to another location.

The dimensional analysis can be used to determine the kind of interaction the slope wind will have with the ambient wind. The nature of the slope wind-ambient wind interaction has been discussed in several papers. The observations presented therein substantiate the persistence of slope wind in the presence of ambient wind. Interpretation, and use of equation 1.6.14 is by an attempt to provide a quick method for estimating whether a slope wind will persist and be observable in the presence of cocurrent or counter current ambient wind.

The following guidelines are proposed when interpreting the value of \tilde{U} . For $|\tilde{U}| \gg 1$ the effects of the slope wind will not be observed. Under such circumstances the atmospheric conditions that produce slope wind are not likely to be maintained.

When $|\vec{U}| \approx 1$ the effect of the slope wind may be evident as a distortion in the logarithmic groundlevel wind velocity profile, but a true jet will not be observed. When $1 > |\vec{U}| > 0$ the observed wind profile will have a low level jet, but also show the influence of the ambient wind with increased vertical height. Naturally when $\vec{U} = 0$ ambient winds are not present and the low level jet will be observed undisturbed. Because observations available from the literature are insufficient to substantiate these criteria, the above are simply guidelines, and may need to be revised as more data becomes available.

Slope winds are often nested within other mesoscale meteorological regimes, for example, the mountain-valley system. Thus it may be possible to obtain estimates of the ambient wind potential in an area also susceptible to slope wind, by techniques developed for these other mesoscale regimes. This information, plus the pertinent parameters available for the slope wind can be combined and used in Equation 1.6.14 to serve as a predictive tool for estimating the near ground behavior of winds along a slope.

It is useful to place several previous papers into the context of the dimensional analysis just presented. Examination of Prandtl's solution reveals that it is the solution of a slope flow for the dimensionless case of $\lambda \gg 1$. In other words, Prandtl's solution is a realistic solution of the extreme case where entrainment of the ambient air leads to a reversal of the potential temperature gradient atop of the boundary layer. If Ohata and Higuchi (1979) had known this, they probably would not have expected their observations on a snow patch, which had a λ of 0.04 – 0.06 (Table III), to match well with Prandtl's solution. The paper by Rao which suggests that Coriolis effects are negligible except for small slope angles should be amended based on Equation 1.6.10. The statement should also include cases where the slope is long, the ground to ambient air temperature gradient is small, or the Coriolis parameter is large. Specifically, cases where the Rossby number is 1 or smaller.

Conclusions Section 1.7

Five major conclusions can be drawn from the previous section.

1. The von Karman-Pohlhausen technique is successfully applied to slope flows.
2. The velocity profile which is obtained compares well with measurements.
3. The slope flow boundary layer thickness grows slightly in a breeze opposing the direction of its flow, reaching a maximum height when the external flow is fifteen percent the size of the characteristic slope flow speed, $U_{\infty} = 0.15U^*$.
4. The slope flow boundary layer thickness decreases steadily in a breeze flowing in the same direction as the flow.
5. A carefully designed, comprehensive slope flow field study will have to be done to determine the adequacy of the above technique. This field study should consider the following advice:
 - a. Use the results from the dimensional analysis to assess/pick a site for the measurements. A site with a gradual slope on the order of 10 km would be ideal, perhaps a glacier.
 - b. Do simultaneous measurements at many locations along the slope. Make certain the leading edge of the flow can be located by first doing diagnostic measurements.
 - c. Estimate the boundary layer thickness for the slope flow *a priori*, confirm it with diagnostic measurements, and then do extensive measurements from ground level up to this height including a reasonable margin for variations. Definitely find the maximum speed at each point along the slope, since testing the theory depends on knowing it.
 - d. A simultaneous atmospheric tracer release might be considered to parameterize turbulence, mixing, and transport in and of the flow. If real time visualization of the slope flow were being done by reconstructing the data as it were being taken, a very controlled tracer release into the leading edge of the flow would be possible.

Conclusions Section 1.8

Four conclusions can be drawn from the above study of the transient flow.

1. The above method described herein provides a route for investigating systems of coupled PDE's derived from boundary layer analysis. Though approximate, the technique preserves a physical interpretation as to the origin of each term, and does not require numerical techniques to complete the analysis. The next extension is to apply the technique to a more complete description of the problem discussed above.
2. Transient slope flows may be difficult to discern from steady slope flows which have boundary layers which have become independent of slope length.
3. Observations must be made with exceptional care so as to document every pertinent parameter to the flow as described in Section 1.6. Only then will it be possible to determine where along the slope of a given flow the boundary layer should become independent of slope length, and where the observed flow was still experiencing transients.
4. When a high quality field study is done these results should be compared with it, and the existing models of other authors.

Conclusions Section 1.9

The critical slope length, l_c , may be more aptly called the effective slope length for steep slopes, since it really is the maximum segment of the total slope length which must be experiencing a potential temperature difference in order to create a flow. For steep slopes, the place along the slope where the katabatic wind develops may be of little importance because the critical slope length is short compared to the total slope length. For the slopes which surround the Los Angeles basin the critical length is modest compared to the total slope lengths, 20 percent or less in all cases studied. In terms of the computation interval of numerical models this means energy and momentum calculations along slopes should be done at intervals of 0.5 km at the most. The small temperature differences which drive the flow, on the order of a degree, and the vertical bound on the slope flow thickness of 100

meters means vertical computational resolution on the order of 30 meters or less. It is clear that considerable computational power is required to reproduce these flows accurately above any large surfaces.

Though no documented evidence is known by the author to support this claim, it is suggested that the intrusion of a large amount of stable slope air centering around Glendora and Azusa may on certain mornings initially slow the transport of western basin emissions eastward. Furthermore, the emissions from the late afternoon and evening of the previous day may be revisiting the basin by means of the slope flow drainage through Azusa and Glendora. The detailed physics of the flow from this area could be studied quite effectively using atmospheric tracers. The knowledge obtained would benefit both the conceptual model and the numerical calculations at this, and other locations.

Conclusions Section 2.1

- (1) In both summer studies, during the hour following the end of a three-hour release, the tracer was dispersed over an area of approximate 400 square kilometers.
- (2) During the summer, an effective eddy diffusivity of around 10_4 meters²/second characterized the dispersion of the tracer.
- (3) Mass balances of the tracer indicated that essentially off of the tracer was accounted for all of the tracer.
- (4) An objective trajectory model was developed to help interpret the tracer data.
- (5) Both the surface and upper level winds are needed to explain the transport of the tracer.
- (6) The morning downtown emissions did not impact the ozone in the eastern

region of the Los Angeles region.

- (7) In both winter studies, a vast majority of the tracer was carried over. The residence time within the SCAQS area of interest was greater than 36 hours.
- (8) Mass balances indicated that a more complex pattern of dispersion occurred during the winter as compared to that during the summer.
- (9) The convoluted surface wind trajectories help verify the presence of flow reversals.
- (10) In both cases the tracer was transported outside of the boundary for which adequate meteorological data are available.
- (11) Increased meteorological data, included (i) winds over the ocean, (ii) winds in the mountains, and (iii) upper levels winds are needed to adequately reconstruct all of the transport and dispersion patterns.
- (12) During December 11, 1987, the tracer was observed to coincide with the elevated carbon monoxide levels at Lynwood.

Conclusions Section 2.3

The six sets of simulations and compilations performed using the 1987 and 1988 wind and ozone data have elucidated the dynamic nature of the LA-Glendale divergence zone and established its persistence. Simulations 4 and 5 clarify that the transport observed during the two summer SCAQS studies were not anomalous observations, but in fact, the norm for the smog season. The comparison of ozone data with the location of the zone in simulation 6 suggests a link between the ozone peaks around sites 1 and 8 and the eastern edge of the zone. The physics of the area around sites 1 and 8 has not been adequately studied so as to clarify the

interaction between the divergence zone, the mountains, ozone precursors, and any other unknown factors.

This more complete understanding of the dynamics of the LA-Glendale divergence zone is likely to improve the overall understanding of the source-receptor relationship that exists between the western basin and the receptors to the east and northwest. Finally the exact location of this zone can be used as a sensitive diagnostic for confirming wind field models, because any model that does not accurately position the LA-Glendale zone is not correctly capturing the physics of the western Los Angeles Basin.

Conclusions Section 2.4

The Cross Differencing Method

1. The cross differencing method is easy to implement, requires commonly available computing hardware, and is fast to use once reading the plots is mastered.
2. The method has been successfully demonstrated using wind fields from both seasonal average days and individual days.
3. A natural application of the method, but one not examined in this paper, is the comparison of two wind fields for the same site. This approach is likely to be very useful in visually identifying times when numerical models succeed and fail at a particular location. Furthermore, the comparison of the results of different numerical models can be accomplished by this method.

Transition Periods in the Los Angeles Basin

1. The transition periods primarily have an effect on the morning rush hour. On average, strong direction change occurs within the morning rush hour at 27 of the 31 sites, while only 2 of the 31 sites (Burbank, Reseda) have direction changes during the evening rush hour.
2. On average, the transitions are characterized by two sharp directional changes;

one occurring at the end of the previous days wind flow and the other at the start of the new days flow. The wind direction often changes on the order of 100 or more degrees on transition.

3. Transport during a one to five hour transition period may relocate emissions from between 3 - 15 kilometers from the source because of the directional nature of the flows during these periods.

SCAQS SF_6 Days

1. The two days examined show similar dynamics both near the downtown Los Angeles release site and at two sites situated near basin air flow exits. Both days had strongly suppressed wind speeds near the downtown area and higher than average speeds at the exits.

Accountability Modeling needs to accurately account for the timing of the transitions, and their flow direction and speed. Mesoscale effects are observed primarily after dark, and most commonly in the early morning hours.

Conclusions Section 3

Neural networks were capable of being configured and trained for the two classes of problems discussed. In both cases the network's performance was adequate for the task being performed, though direct comparisons with other methods were lacking as a definitive yardstick for comparison. Though it was not elaborated on, neuro-computing is the only method known to this author where the prediction or reconstruction made by the network can also include a measure of the networks confidence in its result.

The techniques employed can be extended to more complex cases with more diverse input sets. For instance, the temperature dependence, wind speed, atmospheric mixing depth, and ozone levels at other locations can all serve as inputs to either the data reconstruction or forecasting network. From the previous discussion

it is also clear how networks might be trained to generate wind field forecasts at least several hours into the future. Such a tool might be useful to meteorologists and AQ specialists.

Appendix B.1 - Dynamically Allocated Memory

Runtime Memory Requirements

WIND2D	125,424 bytes
WINND3D	4,116,536 bytes

Due to current technical limitations the PC is only comfortable at manipulating 640,000 bytes in its internal memory. From the above information it is possible to see why WIND2D was easily converted, while WIND3D was not.

ASPECTS OF ATMOSPHERIC TRANSPORT AND
DISPERSION WITHIN AN AIR BASIN

Thesis by
Robin Scott Horrell

In Partial Fulfillment of the Requirements
for the Degree of
Doctor of Philosophy

California Institute of Technology
Pasadena, California

1991
(Submitted January 10, 1991)

©1991

Robin Scott Horrell

All Rights Reserved

This thesis is dedicated

to

my parents and
Andrea, my lovely wife

Acknowledgements

It is impossible to thank everyone for every contribution which has ultimately been manifested in the final dissertation. To this end, I would first like to thank everyone from whom I received help and suggestions and which I might have inadvertently omitted in the following acknowledgements.

I would like to thank the Air Quality Management District for access to equipment, and extensive access to personnel. For access to their computer systems and telemetry system I would like to thank Ron Sweet and Chris Marlia. The help of the meteorologists Bruce Selik and Jerry Arnold especially during the South Coast Air Quality Study of 1987 was highly appreciated. Within the planning division I would especially like to thank Chung Liu, Henry Hogo, and Tom Chico for their helpful suggestions and assistance during my research. I wish to give special thanks to Julia Lester, also of the planning division, for her suggestions, corrections, and constructive criticisms of many sections of this thesis.

I owe a special thank you to Joe Cassmassi, the AQMD's Senior Meteorologist. His suggestions in the development of the WIND simulation package were numerous and constructive. On a personal level I would also like to acknowledge Joe for his genuine interest in my professional development. I admire his ability to integrate people and in my case, in doing so, make me strive to bring out my best. Interaction with Joe will have a lasting influence on my measure of what excellence is in the work place.

The California Air Resources Board (CARB) has supported a large fraction of this research. Within the CARB I would like to thank Bart Croes, Lowel Ashbaugh and especially Chuck Bennett for favors too numerous to list. I would also like to thank Don Lehrman of Technical and Business Systems for supplying me with preliminary upper air data from the SCAQS.

At Caltech I owe special thanks to Mike Deem who has been both an assis-

tant, friend, and inspiration. His conceptualization of the WIND program from my original coarse formulation is exceptional. Peter Wyckoff and Scott Halliday also at Caltech have contributed significantly to the development and coding of many of my computer programs. I would like to thank four people whose friendship made my time at Caltech more enjoyable, Chris Guske, Greg Sullivan, Ron Benson and Mike Loewenberg.

A special thanks is due to my wife, Andrea, who made suggestions, corrections, and overall endured me at times when I could hardly endure myself. Her efforts and tolerance will long be remembered and appreciated.

Finally, I would like to thank my advisor Dr. Fred Shair who has patiently taught me a great deal about science and engineering. Without his help and guidance, none of this would have been possible. I would like to extend that thanks to him for his generosity with his time and his willingness to allow me to explore subjects which did not directly contribute to this thesis but greatly contributed to my education.

Thank you.

Abstract

Atmospheric transport and dispersion relating to observed pollution levels are analyzed using three methods. Two traditional methods, (1) dimensional analysis and (2) atmospheric sulfur hexafluoride tracer experiments are applied to new problems; the third, (3) neural computation is a new method of looking at incomplete ozone data reconstruction, and short term ozone forecasting.

A comprehensive dimensional analysis of buoyantly driven slope wind is presented to unify the subject. The relationship between thickness, velocity, Reynolds, Froude, Rossby and Richardson numbers of the buoyant layer and the characteristic slope parameters is established. A detailed study of the velocity and temperature profiles is done using the Von Karman-Pohlhausen (KP) integral technique. Generated profiles are in agreement with field measurements. The technique is extended for transient slope flows using a combination of the KP method and the method of characteristics.

The transport and dispersion of airborne emissions from Los Angeles are studied using tracer, during the Southern California Air Quality Study of 1987. Emphasis is placed on emissions generated during the morning land breeze/sea breeze transition periods. Analysis of tracer trajectories indicates emission transport from downtown Los Angeles was by two routes, and the relative importance of each route was determined by a diurnally occurring divergence zone. This zone, resulting from geography and pressure gradients, was shown to migrate east to west in the air space above Los Angeles. The zone routes a majority of morning emissions northwest of Los Angeles, contrary to the commonly held belief that morning emissions were routed east. Observations were verified using surface wind trajectories constructed from telemetry data.

Neural computational strategies are applied to two ozone data sets for the purpose of data reconstruction and forecasting at Newhall, California. The Madaline III algorithm is used to reconstruct data at a single site. The reconstruction is done to add missing data. This method is advantageous when

peaks in the data are missing, and traditional interpolation methods inapplicable. Williams and Zipser's recurrent algorithm is used to make short-term ozone forecasts at a single location based on past history of ozone observations at that site. The results are systematically better than regression methods.

Table of Contents

	Page
Copyright Page	ii
Dedication	iii
Acknowledgements	iv
Abstract	vi
Table of Contents	viii
List of Figures	x
List of Tables	xxii
Introduction	1
Chapter One - A Boundary Layer Analysis of Bouyantly Driven Slope Flow	
A Comprehensive Dimensional Analysis for Assessing the Transport Properties of Atmospheric Slope Flows (Sections 1.1-1.6)	
1.1 Introduction	8
1.2 Theory	20
1.3 Making the Continuity Equation Dimensionless	23
1.4 Making the Energy Equation Dimensionless	27
1.5 Making the Momentum Equation Dimensionless	31
1.6 Characeristic and Dimensionless Slope Flow Parameters	36
1.7 Application of the van Karman-Pohlhausen Method to a Bouyantly Driven Boundary Layer	51
1.8 Modeling the Evolution of a Two-Dimensional Thermally Driven Atmospheric Boundary Layer	67
1.9 Katabatic Wind Along the Mountain Slopes Bordering the Los Angeles Basin	75
A1.9 Appendix	93
References	110

Chapter Two - A Study of Atmospheric Transport and Dispersion Near a Two-Dimensional Divergence Zone	
2.1 Ground Release SF_6 Tracer Experiments Used to Characterize Transport and Dispersion of Atmospheric Pollutants During the Southern California Air Quality Study of 1987	113
2.2 Establishing Upper Bounds on Vertical SF_6 Tracer Transport During the 1987 SCAQS	140
A2.2 Appendix	155
References	192
2.3 A Smog Season Examination of the LA-Glendale Divergence Zone	193
A2.3 Appendix	229
References	232
2.4 Examination of Wind Field Dynamics Using Cross Differencing	234
A2.4 Appendix	276
References	293
Chapter Three - Neural Network Computing Techniques to Process Air Quality Data	
3.1 Introduction	296
3.2 Neuro-computing Strategy	298
3.3 A Feedforward Network for Ozone Data Reconstruction	304
3.4 A Recurrent Network for Short-term Ozone Forecasting	310
References	323

List of Figures

Chapter One

Section 1.1

1. An example of the geography giving rise to slope flow. 9
2. An example of the potential temperature and velocity profiles in a slope flow. 10

Section 1.2

1. Slope coordinates relative to the rectangular coordinates of level ground. 24

Section 1.6

1. Linear correlation between the observed maximum velocity, U_{Obs}^{Max} , of the low level jet and the characteristic velocity, U^* , obtained via dimensional analysis. 42

Section 1.7

1. Comparison of the velocity profile given by Equation 1.7.7. 61
2. Changes in the slope flow momentum boundary layer thickness in the presence of an external constant flow. 63

Section 1.8

1. Characteristic lines for general transient slope flow. 72
2. Boundary layer thickness as a function of dimensionless slope length and dimensionless time. 73

Section 1.9

1. Velocity and temperature profiles in an idealized slope flow. The boundary layer thickness, Δ , the slope angle, ϕ , and slope length, L , are labeled. 77
2. Slope wind measuring sites in the Los Angeles Basin. 78
3. Characteristic slope wind velocity as a function of slope 88

length and slope strength also showing lines of critical slope length for different slope angles.

A 1.9 Appendix

1. Northwestern basin sites in the San Fernando Valley.	100
2. Northern basin sites bordering mountains.	101
3. North Central site bordering mountains with a strong nighttime slope flow.	102
4. Eastern basin sites (1).	103
5. Eastern basin sites (2).	104
6. Eastern desert sites.	105
7. Southwest coastal regions.	106
8. Central basin sites.	107
9. Downtown and western coastline sites.	108
10. Sites on basin's perimeter.	109

Chapter Two

Section 2.1

1. Location of reporting sites for the SCAQMD Telemetry System in the western Los Angeles Basin.	130
2. Location of upper air network stations, and additional one hour average sample board locations.	131
3. SF_6 tracer cloud as seen by auto traverses on July 15, 1987, (a) 9:00 - 9:30 PDT and (b) 9:30 - 10:30 PDT.	132
4. SF_6 tracer cloud as seen by auto traverses on September 3, 1987, (a) 9:00 - 10:00 PDT and (b) 13:00 - 14:00 PDT.	133
5. Surface wind trajectories from Vernon and other locations at 6:00, 7:00, and 8:00 PDT, and lasting 24 hours (a) July 15, 1987, (b) September 3, 1987, and upper air trajectories starting at 6:00 PDT for altitudes of	134

100, 200, 300, and 400 meters AGL going counter-clockwise about Vernon (c) July 15, 1987, and (d) September 3, 1987.	
6. SF_6 tracer cloud as seen by auto traverses on (a) at 6:00 - 8:00 PST November 13, 1987, 24 hours after the release, (b) at 19:00 PST December 10, 1987, just after the evening release.	135
7. SF_6 tracer cloud as seen by auto traverses on December 11, 1987, the day after the release (a) at 7:35 PST, (b) at 14:00 PST.	136
8. Surface wind trajectories from Vernon starting at (a) 6:00, 7:00, and 8:00 PST, lasting 36 hours on November 12-13, 1987, and (b) 16:00, 17:00, and 18:00 PST lasting 24 hours on December 10-11, 1987.	137
9. Upper air wind trajectories from Vernon starting on November 12, 1987, starting at 6:00 PST and lasting 18 hours at altitudes AGL of (a) 100 meters, (b) 200 meters, (c) 300 meters, (d) 400 meters.	138
10. Upper air wind trajectories from Vernon starting on December 10, 1987, starting at 16:00 PST and lasting 18 hours at altitudes AGL of (a) 200 meters, (b) 300 meters, (c) 400 meters, (d) 600, 800, 1000 and 2000 meters going clockwise about Vernon.	139

Section 2.2

1. Idealized vertical temperature profiles.	145
2. Slope of the potential temperature with increasing height versus range of atmospheric stability.	150

A2.2 Appendix

1. Vertical atmospheric structure at Burbank at 05:00 PDT, July 15, 1987.	156
2. Vertical atmospheric structure at Burbank at 08:00 PDT, July 15, 1987.	156
3. Vertical atmospheric structure at Burbank at 11:00 PDT,	157

July 15, 1987.

4. Vertical atmospheric structure at Burbank at 15:00 PDT, July 15, 1987.	157
5. Vertical atmospheric structure at Burbank at 17:00 PDT, July 15, 1987.	158
6. Vertical atmospheric structure at Burbank at 22:00 PDT, July 15, 1987.	158
7. Vertical atmospheric structure at El Monte at 05:00 PDT, July 15, 1987.	159
8. Vertical atmospheric structure at El Monte at 08:00 PDT, July 15, 1987.	159
9. Vertical atmospheric structure at El Monte at 11:00 PDT, July 15, 1987.	160
10. Vertical atmospheric structure at El Monte at 14:00 PDT, July 15, 1987.	160
11. Vertical atmospheric structure at El Monte at 18:00 PDT, July 15, 1987.	161
12. Vertical atmospheric structure at Burbank at 05:00 PDT, September 3, 1987.	162
13. Vertical atmospheric structure at Burbank at 08:00 PDT, September 3, 1987.	162
14. Vertical atmospheric structure at Burbank at 11:00 PDT, September 3, 1987.	163
15. Vertical atmospheric structure at Burbank at 14:00 PDT, September 3, 1987.	163
16. Vertical atmospheric structure at Burbank at 17:00 PDT, September 3, 1987.	164
17. Vertical atmospheric structure at Burbank at 22:00 PDT, September 3, 1987.	164
18. Vertical atmospheric structure at El Monte at 08:00 PDT, September 3, 1987.	165

19. Vertical atmospheric structure at El Monte at 12:00 PDT, September 3, 1987.	165
20. Vertical atmospheric structure at El Monte at 14:00 PDT, September 3, 1987.	166
21. Vertical atmospheric structure at El Monte at 17:00 PDT, September 3, 1987.	166
22. Vertical atmospheric structure at Burbank at 04:00 PDT, November 12, 1987.	167
23. Vertical atmospheric structure at Burbank at 08:00 PDT, November 12, 1987.	167
24. Vertical atmospheric structure at Burbank at 11:00 PDT, November 12, 1987.	168
25. Vertical atmospheric structure at Burbank at 14:00 PDT, November 12, 1987.	168
26. Vertical atmospheric structure at Burbank at 17:00 PDT, November 12, 1987.	169
27. Vertical atmospheric structure at Burbank at 22:00 PDT, November 12, 1987.	169
28. Vertical atmospheric structure at El Monte at 06:00 PDT, November 12, 1987.	170
29. Vertical atmospheric structure at El Monte at 11:00 PDT, November 12, 1987.	170
30. Vertical atmospheric structure at El Monte at 14:00 PDT, November 12, 1987.	171
31. Vertical atmospheric structure at El Monte at 17:00 PDT, November 12, 1987.	171
32. Vertical atmospheric structure at Long Beach at 05:00 PDT, November 12, 1987.	172
33. Vertical atmospheric structure at Long Beach at 08:00 PDT, November 12, 1987.	172
34. Vertical atmospheric structure at Long Beach at 11:00 PDT,	173

November 12, 1987.

35.	Vertical atmospheric structure at Long Beach at 14:00 PDT, November 12, 1987.	173
36.	Vertical atmospheric structure at Long Beach at 17:00 PDT, November 12, 1987.	174
37.	Vertical atmospheric structure at Long Beach at 22:00 PDT, November 12, 1987.	174
38.	Vertical atmospheric structure at Loyola Marymount University at 05:00 PDT, November 12, 1987.	175
39.	Vertical atmospheric structure at Burbank at 11:00 PDT, December 10, 1987.	176
40.	Vertical atmospheric structure at Burbank at 14:00 PDT, December 10, 1987.	176
41.	Vertical atmospheric structure at Burbank at 17:00 PDT, December 10, 1987.	177
42.	Vertical atmospheric structure at Burbank at 22:00 PDT, December 10, 1987.	177
43.	Vertical atmospheric structure at El Monte at 05:00 PDT, December 10, 1987.	178
44.	Vertical atmospheric structure at El Monte at 08:00 PDT, December 10, 1987.	178
45.	Vertical atmospheric structure at El Monte at 11:00 PDT, December 10, 1987.	179
46.	Vertical atmospheric structure at El Monte at 15:00 PDT, December 10, 1987.	179
47.	Vertical atmospheric structure at Long Beach at 11:00 PDT, December 10, 1987.	180
48.	Vertical atmospheric structure at Long Beach at 14:00 PDT, December 10, 1987.	180
49.	Vertical atmospheric structure at Long Beach at 17:00 PDT, December 10, 1987.	181

50. Vertical atmospheric structure at Long Beach at 22:00 PDT, December 10, 1987.	181
51. Vertical atmospheric structure at Loyola Marymount University at 00:00 PDT, December 10, 1987.	182
52. Vertical atmospheric structure at Loyola Marymount University at 14:00 PDT, December 10, 1987.	182
53. Vertical atmospheric structure at Loyola Marymount University at 17:00 PDT, December 10, 1987.	183
54. Vertical atmospheric structure at Burbank at 05:00 PDT, December 11, 1987.	184
55. Vertical atmospheric structure at Burbank at 08:00 PDT, December 11, 1987.	184
56. Vertical atmospheric structure at Burbank at 10:00 PDT, December 11, 1987.	185
57. Vertical atmospheric structure at Burbank at 14:00 PDT, December 11, 1987.	185
58. Vertical atmospheric structure at El Monte at 08:00 PDT, December 11, 1987.	186
59. Vertical atmospheric structure at El Monte at 11:00 PDT, December 11, 1987.	186
60. Vertical atmospheric structure at El Monte at 17:00 PDT, December 11, 1987.	187
61. Vertical atmospheric structure at Long Beach at 05:00 PDT, December 11, 1987.	188
62. Vertical atmospheric structure at Long Beach at 08:00 PDT, December 11, 1987.	188
63. Vertical atmospheric structure at Long Beach at 11:00 PDT, December 11, 1987.	189
64. Vertical atmospheric structure at Long Beach at 14:00 PDT, December 11, 1987.	189
65. Vertical atmospheric structure at Loyola Marymount University	190

at 05:00 PDT, December 11, 1987.

66. Vertical atmospheric structure at Loyola Marymount University at 08:00 PDT, December 11, 1987.	190
67. Vertical atmospheric structure at Loyola Marymount University at 11:00 PDT, December 11, 1987.	191
68. Vertical atmospheric structure at Loyola Marymount University at 14:00 PDT, December 11, 1987.	191

Section 2.3

1. Wind vectors at 8 a.m. for the 1988 composite smog season day.	200
2. Wind vectors at 3 p.m. for the 1988 composite smog season day.	208
3. Angular divergence map at 8 a.m. for the 1988 composite smog season day.	209
4. Angular divergence map at 3 p.m. for the 1988 composite smog season day.	210
5. Simulated 12 hour tracer releases at 6, 7, and 8 a.m. for 4 sites (a) West LA (b) downtown LA (c) Hawthorne (d) Lynwood.	214
6. Simulated 12 hour tracer releases at 6, 7, and 8 a.m. for 5 sites (a) Pico Rivera (b) Whittier (c) southwest coastline and Long Beach (d) Los Alamitos.	215
7. Simulated 12 hour tracer releases at 3, 4, and 5 p.m. for 4 sites (a) West LA (b) downtown LA (c) Hawthorne (d) Lynwood.	216
8. Simulated 12 hour tracer releases at 3, 4, and 5 a.m. for 5 sites (a) Pico Rivera (b) Los Alamitos and Whittier (c) southwest coastline (d) Long Beach.	217
9. Percentage of western basin trajectories exiting via the eastern basin from 4 sites: West LA, downtown LA, Hawthorne, and Lynwood.	219

10. Range of motion of the LA-Glendale divergence zone.	222
11. 1988 smog season ozone level contour map at 8 a.m.	225
12. 1988 smog season ozone level contour map at 3 p.m.	226

A2.3 Appendix

A1. Distribution of mobile source emissions of reactive organic gases.	230
A2. Spatial distribution of population density.	231

Section 2.4

1. Hourly averaged wind vectors for the 1988 composite average day during the smog season at 8 a.m.	236
2. Cross differencing example 1.	245
3. Cross differencing example 2.	249
4. Cross differencing example 3.	251
5. Cross differencing example 4.	254
6. Cross differencing example 5.	256
7. Cross differencing example 6.	257
8. Cross differencing example 7.	259
9. Speed and angle cross difference at downtown LA (CELA) (top) July 15, 1987, (bottom) September 3, 1987.	270
10. Speed and angle cross difference at Burbank (BURK) (top) July 15, 1987, (bottom) September 3, 1987.	271
11. Speed and angle cross difference at Upland (UPLA) (top) July 15, 1987, (bottom) September 3, 1987.	272

A2.4 Appendix

12. Speed and angle cross difference (top) Anaheim (ANAH) average smog season day 1988 compared to itself, (bottom)	277
---	-----

Azusa (AZUS) average smog season day 1988 compared to itself.	
13. Speed and angle cross difference (top) Banning (BANN) average smog season day 1988 compared to itself, (bottom) Burbank (BURK) average smog season day 1988 compared to itself.	278
14. Speed and angle cross difference (top) Central Los Angeles (CELA) average smog season day 1988 compared to itself, (bottom) Costa Mesa (COST) average smog season day 1988 compared to itself.	279
15. Speed and angle cross difference (top) Crestline (CRES) average smog season day 1988 compared to itself, (bottom) Fontana (FONT) average smog season day 1988 compared to itself.	280
16. Speed and angle cross difference (top) Glendora (GLEN) average smog season day 1988 compared to itself, (bottom) Hawthorne (HAWT) average smog season day 1988 compared to itself.	281
17. Speed and angle cross difference (top) Hemet (HEME) average smog season day 1988 compared to itself, (bottom) La Habra (LAHB) average smog season day 1988 compared to itself.	282
18. Speed and angle cross difference (top) Lancaster average smog season day 1988 compared to itself, (bottom) Long Beach (LGBH) average smog season day 1988 compared to itself.	283
19. Speed and angle cross difference (top) Los Alamitos (LSAL) average smog season day 1988 compared to itself, (bottom) Lynwood (LYNN) average smog season day compared to itself.	284
20. Speed and angle cross difference (top) Newhall (NEWL) average smog season day 1988 compared to itself, (bottom) Norco (NORC) average smog season day compared to itself.	285
21. Speed and angle cross difference (top) Pasadena (PASA) average smog season day 1988 compared to itself, (bottom) Perris (PERI) average smog season day compared to itself.	286

22. Speed and angle cross difference (top) Pico Rivera (PICO) average smog season day 1988 compared to itself, (bottom) Palm Springs (PLSP) average smog season day compared to itself.	287
23. Speed and angle cross difference (top) Pomona (POMA) average smog season day 1988 compared to itself, (bottom) Redlands (RDLD) average smog season day compared to itself.	288
24. Speed and angle cross difference (top) Reseda (RESE) average smog season day 1988 compared to itself, (bottom) Riverside (RIVR) average smog season day compared to itself.	289
25. Speed and angle cross difference (top) San Bernardino (SNBO) average smog season day 1988 compared to itself, (bottom) El Toro (TORO) average smog season day compared to itself.	290
26. Speed and angle cross difference (top) Upland (UPLA) average smog season day 1988 compared to itself, (bottom) Whittier (WHIT) average smog season day compared to itself.	291
27. Speed and angle cross difference (top) West Los Angeles (WSLA) average smog season day 1988 compared to itself.	292

Chapter Three

Section 3.3

1. Input, output, target and error for the sample input July 14, 1989: (top) complete input, (bottom) input with two deletions at hours 13 and 14.	306
2. Input, output, target and error for the sample input July 6, 1989: (top) complete input, (bottom) input with two deletions at hours 12 and 16.	308

Section 3.4

1. Average sum square error versus the number of neurons.	312
2. Test data: ozone forecast error versus the hour of the day.	313

3. Training data: ozone forecast error versus the hour of the day.	314
4. Ozone forecast versus the actual observed level three hours ahead.	316
5. Recurrent network output at start-up for predictions four, five, and six hours into the future.	317
6. Output of the three hidden neurons as a function of time when predicting four, five, and six hours into the future.	318
7. Comparison of three ozone forecast methods: (1) Recurrent network, (2) Linear regression, and (3) Seasonal Average.	320

List of Tables

Chapter One

Section 1.6

- | | |
|---|----|
| 1. Observations of buoyantly driven flows:
locations, sites, and authors. | 39 |
| 2. Observations of buoyantly driven flows:
critical parameters, characteristic and observed velocities. | 40 |
| 3. Observations of buoyantly driven flows:
characteristic time, critical slope length ratio (λ),
and Rossby number. | 43 |

Section 1.9

- | | |
|---|----|
| 1. Observation sites, and nearby slope's angle and length. | 81 |
| 2. Nighttime slope/drainage wind characteristics. | 81 |
| 3. Nighttime differential temperature, critical
slope length, and dimensionless slope parameter. | 90 |

A1.9 Appendix

- | | |
|--|----|
| 1. Average wind speed from 31 SCAQMD Telemetry
System Sites for the period May-September 1988 at all 24 hours. | 96 |
| 2. Standard deviation of the wind speed from
31 SCAQMD Telemetry System Sites for the period
May-September 1988 at all 24 hours. | 97 |
| 3. Average wind direction from 31 SCAQMD
Telemetry System Sites for the period May-September 1988
at all 24 hours. | 98 |
| 4. Standard deviation of the wind direction from
31 SCAQMD Telemetry System Sites for the period
May-September 1988 at all 24 hours. | 99 |

Chapter Two

Section 2.1

1. Station code and name for SCAQMD Telemetry Stations.	127
2. Station code and name for SCAQS upper air stations.	127
3. Code and name for additional automatic sampler locations.	127
4. a. Automatic sampler results, November 12.	128
4. b. November 13.	128
5. a. Automatic sampler results, December 10.	129
5. b. December 11.	129
5. c. December 112 (continued).	129

Section 2.2

1. Mixing depths during the summertime SCAQS 1987.	152
2. Mixing depths during the November SCAQS 1987.	153
3. Mixing depths during the December SCAQS 1987.	154

Section 2.3

1. Air quality and meteorological reporting stations.	202
2. Percentage of days where the LA-Glendale divergence zone is identifiable at a particular time.	203
3. Release points in the Western Los Angeles Basin.	213

Section 2.4

1. Air quality and meteorological reporting stations.	246
2. Example 24 element sequences which are used to create a number of possible features in the cross differencing plots in Appendix 1.	247
3. Features of the average angle difference plots at 31 sites.	265

Introduction

This thesis is divided into three chapters, and each chapter is divided into sections. The tables and figures of each section are numbered independently. Each equation is numbered by chapter, section, then sequentially within the section, e.g., Equation 1.2.7 refers to Chapter One, Section 2, Equation 7. The contents of the chapters are summarized as follows. Chapter One is a dimensional analysis of atmospheric slope wind including several applications of the resulting analysis. Chapter Two is a study of atmospheric transport and dispersion near a two-dimensional divergence zone. Chapter Three is a discussion of the application of neuro-computing strategies to air quality modeling, and to processing air quality data.

The Sections 1.1 and 1.2 of Chapter One introduce the subject of buoyantly driven slope wind. In Sections 1.3-1.5 a boundary layer analysis of this subject is performed. In Section 1.6 the characteristic and dimensionless parameters from the boundary layer analysis are summarized and compared to worldwide observations. In Sections 1.7, the approximate boundary layer technique devised by von Karman and Pohlhausen is applied to the results of Sections 1.4 and 1.5 for steady state slope flows. Section 1.8 extends the analysis of Section 1.7 to include transient slope flows. Finally, in Section 1.9 ten sites in the Los Angeles Basin which exhibit slope flows are analyzed using the above framework. This analysis makes clear how the mesoscale slope flows associated with the mountains surrounding a basin can be transporting natural and anthropogenic emissions both into and out of the basin by upslope and downslope winds.

While reading Chapter One keep in mind a few conventions used throughout the text. One, slope winds are often difficult to observe during the day because they are obscured by other stronger winds, e.g., a sea breeze. Thus, when observations of slope winds are discussed, they are in general downslope wind and may be referred to only as *slope wind*. Two, a general word *katabatic* exists, meaning of a nighttime wind which gravitates downslope due to a density differ-

ence between surface and ambient air. Katabatic is used interchangeably with downslope as a descriptor. Three, the terminology *slope wind* and *slope flow* are also synonymous.

Chapter Two addresses the general problem how to assess the atmospheric transport and dispersion within an air basin in which the multiple inlets and outlets create convergence and divergence zones. The Los Angeles-Glendale divergence zone of Los Angeles Basin is studied by atmospheric tracer experiments and the use of surface winds. In this case the incoming westerly and southerly sea breeze is divided by the divergence zone in the vicinity of downtown Los Angeles. The emissions put into the breeze prior to the branch point of the divergence zone are split and routed through the basin through two separate and distant exits. Chapter Two Section 2.1 describes four sulfur hexafluoride tracer studies done in the proximity of the branch point as part of the Southern California Air Quality Study (SCAQS) of 1987. A part of Section 2.1 is devoted to performing mass balances on the tracer, and Section 2.2 is a detailed description of the measurements of the vertical atmospheric structure required to close the mass balances. Section 2.3 places the observations made in the two summer sulfur hexafluoride studies into the context of the air flow patterns from an entire summer season. Finally, Section 2.4 describes a new, alternative method for examining the air flow dynamics at a single site, by comparing the dynamics at that site to a well chosen reference. This method is applied to the wind fields near downtown Los Angeles on the two summertime tracer days, and to summer season averages from all of the South Coast Air Quality Managements Districts Air Monitoring sites.

An implicit theme of Chapter Two is visual and interactive automation of meteorological and air quality data processing. Each section of the analysis is assisted by WIND, a PC-based collection of interactive programs to process meteorological (MET) and air quality (AQ) data, written to accomplish the analyses in Chapter Two. The philosophy that guided the entire development of

WIND was the ability for the end user of the MET and AQ data to be able to (1) view the data graphically in nearly real time, (2) view MET and AQ data simultaneously, (3) interactively perform a large number of tasks which are desirable to repeat, such as simulating air parcel trajectories, conducting mass balances, and determining fluxes at any location. In addition, the philosophy was adopted that the program should be capable of being used in any geographic location where data was available, and be written in well structured C code so as to be portable to other computer environments if at some later time this became desirable. WIND programs are briefly discussed in each section where they are used, and are documented by a separate user's guide, currently version 2.0, and the separate technical guide of C language source code.

Chapter Three begins with a motivation for examining neuro-computing strategies as they relate to the study of air quality (Section 3.1). Section 3.2 is a summary of how neuro-computers of a particular sort perform their calculation, and are trained. Section 3.3 demonstrates the use of a feedforward network to reconstruct incomplete ozone data and 3.4 demonstrates the use of a recurrent network to do ozone forecasting.

Chapter One

A Boundary Layer Analysis of Buoyantly Driven Slope Flow

Table of Nomenclature

C_p	Heat capacity
g	Gravitational constant, 9.8 ms^{-1}
k	von Karman constant, 0.4
K, K_T	Turbulent transfer coefficients
l	Critical length, either l_c or l_{Co}
L	Characteristic length in the flow direction
n	Dimensional direction normal to the slope
p, p_c	Pressure, characteristic pressure
s	Dimensional direction along the slope
t	Time
t_c	Characteristic time
\bar{t}	Dimensionless time
T	Denotes thermal boundary layer or turbulent, as in eddy viscosity, ν_T
u	Dimensional velocity in the s direction
U	Dimensionless velocity in the slope direction
U^*	Characteristic velocity, b.l. analysis
\bar{U}	Dimensionless ratio of $\frac{U_\infty}{U^*}$
\mathbf{v}	Vector velocity
v	Dimensional velocity in the y direction
V	Dimensionless cross slope velocity
w	Dimensional velocity in the n direction
W	Dimensionless velocity normal to the slope
x	Dimensional distance non-rotated frame
y	Dimensional distance in the width dimension
z	Dimensional dist. in vertical non-rotated frame

α	Thermal diffusivity
β	Coefficient of expansivity, ideal gas $\beta = \frac{1}{T}$
δ	Boundary layer thickness, as a function of χ
f	Coriolis parameter
Δ	Boundary layer vertical characteristic length
Γ	Adiabatic lapse rate
λ	Dimensionless number, effect of lapse rate
Θ	Dimensionless temperature
μ	Viscosity
ν	Kinematic viscosity
ξ	Dimensionless distance along y direction
Ω	Earth's rotation rate
ψ	Latitudinal angle off the equator
ρ	Density
ϕ	Slope angle
τ	Shear stress, both molecular and Reynolds stresses
χ	Dimensionless direction in the slope direction
ζ	Dimensionless direction normal to the slope
∞	Denotes ambient atmospheric conditions
Fr	Froude number
Gr	Grashof number
Pr	Prandtl number
Re	Reynolds number
Ri	Richardson number
Ro	Rossby number
Sc	Schmidt number

**A Comprehensive Dimensional Analysis
for Assessing the Transport Properties of
Atmospheric Slope Flows
(Sections 1.1 - 1.6)**

Abstract

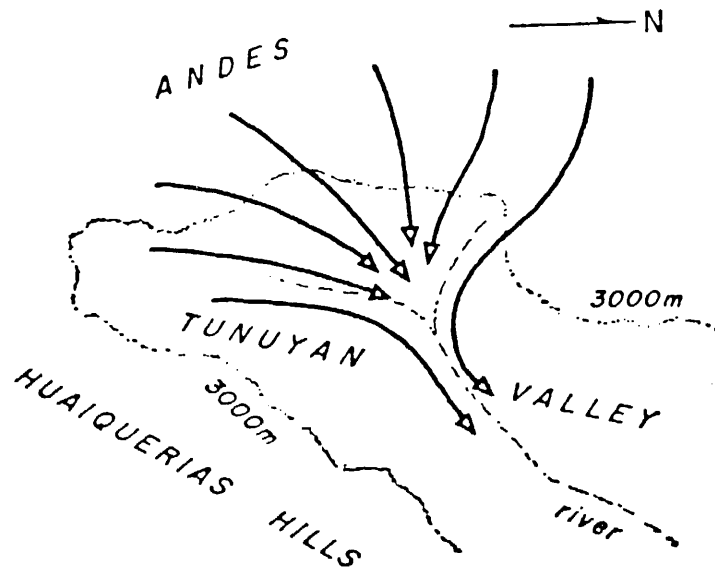
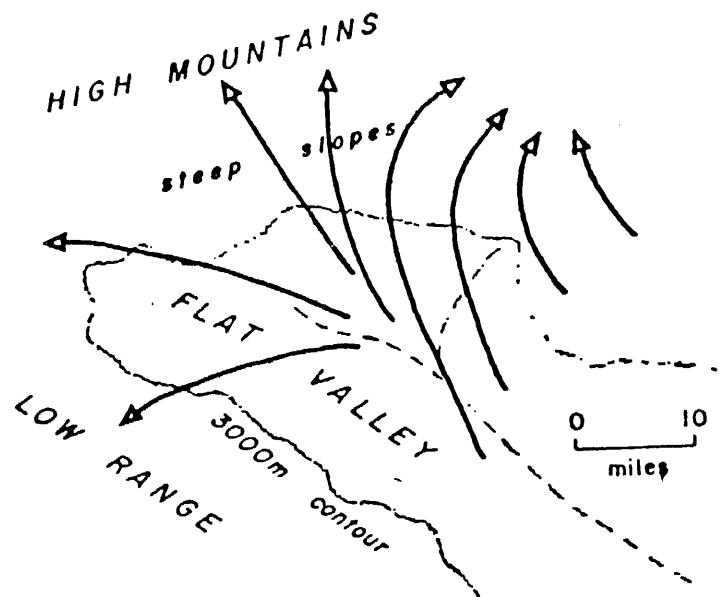
A boundary layer analysis of buoyantly driven atmospheric slope flow is presented. From this analysis a complete set of scaling parameters and dimensional groups are obtained. Among the results from this analysis is a scaling parameter for the heretofore undefined thickness of the buoyant layer. Other simple relationships are derived from this result. These relationships include simple formulations for the characteristic velocity, the Reynolds number, the Rossby number, the Grashof number, and the bulk Richardson number, which can all be expressed in terms of the fundamental parameters of slope length, slope angle, ambient temperature, differential temperature, atmospheric lapse rate, and eddy diffusivity. The characteristic velocity, U^* , is regressed against observations of maximum velocity and is found to follow the relationship $U_{Obs}^{Max} = 0.625U^*$ with a correlation coefficient of 0.913. This suggests that the more complex formulations for maximum velocities in the literature may be unnecessary. In total, the analysis serves to unify the subject of slope flow and to guide investigators at a simple, but fundamental level.

1.1 Introduction

Slope wind is a ubiquitous feature of non-level terrain where a differential temperature between the ground and the ambient air produces a buoyant force that motivates air flow. See Figure 1. Observations show the buoyant force is often large enough in magnitude to oppose the synoptic pressure force and generate and sustain mesoscale air flows along the slope, bearing no similarity to those anticipated by examining synoptic data alone. Buoyantly-driven mesoscale wind, traverse to and cocurrent with the synoptic pressure gradients, can noticeably alter the vertical wind profile near the ground by creating a low level jet. See Figure 2.

Mesoscale slope winds are generally a recurrent, semi-predictable feature of a particular geographic region. Radiative ground cooling, ground heating by insolation, evaporative cooling and the rapid passage of a front, are the most common factors leading to the atmospheric conditions required for buoyantly-driven wind. A particular geographic region and its surroundings may be prone to one or more of the above factors. Consequently, meteorological forecasts that anticipate the occurrence of such factors also preempt the likely observance of mesoscale buoyantly driven wind. The diurnal nature of these factors generally limits the observance of mesoscale slope wind to time scales less than a day, with few exceptions.

Mesoscale buoyantly driven flows are common above and surrounding uneven terrain. Therefore, any serious investigation and interpretation of atmospheric transport in a domain susceptible to the phenomena should not overlook their possible influences. In this respect, three general categories of influence exist for a given geographic region: (1) buoyancy plays little or no role compared to the synoptically driven winds, (2) buoyant effects explain certain anomalous

Mountain wind (night)*Valley wind (day)*

An excellent example of the complete reversal of air flow patterns with the change from day to night is found in the Tunuyan Valley, in the foothills of the Andes. Arrows show how the winds sweep down the cooled slopes into the valley at night but rush up the heated valley slopes during the day.

Figure 1. An example of the geography giving rise to slope flow. (Reproduced from "Watching for the Wind," Edinger, 1967).

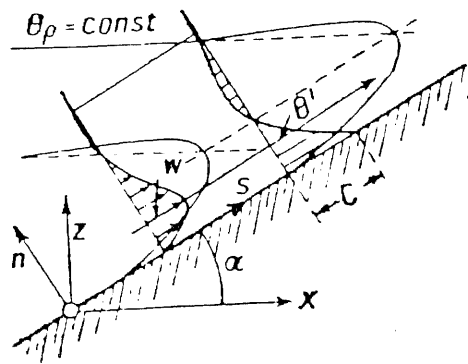


Fig. 5.57.—Wind at a slope

Figure 2. An example of the potential temperature and velocity profiles in a slope flow. (Reproduced from "Essentials of Fluid Dynamics," Prandtl, 1952).

but generally repeatable observations that occur rather unpredictably, and (3) buoyancy accounts for the principle mechanism of local transport. The third category of influence is generally predictable and is only influenced by the passage of very dominant weather systems or seasonal variations.

The work of previous investigators provides insight into the wide range of observations and the present analytical and computational methods used to examine buoyantly driven slope flows. The discussion will be limited to relatively flat geography, i.e., sloping terrain that does not channel or force convergence of the wind as it moves in the flow direction.

Observational Studies

The literature contains many observational studies of atmospheric phenomena dedicated to slope flow, and other studies indirectly present observations pertinent to the subject. Many locations throughout the globe provide persistent, seasonally semi-permanent examples of the slope flow phenomena. The conditions necessary to produce slope wind exist on a wide variety of length, slope and differential (ground to ambient air) temperature scales. The range of these scales is limited only by the physical dimensions and energy budget of the earth. On a continental scale, for instance, the 1000 km slopes with a 0.3 degree angle and small differential temperature of about 0.5 degrees C are responsible for observations made in Antarctica. These produce high velocities, reaching 45 ms^{-1} (around 90 mph), making them the most extreme observation of this phenomena on earth and an interesting subject for study (Ball 1956, 1957, Chiba et al. 1986, Gosink 1987). Other examples on a smaller scale are the slope winds observed on mountain slopes and broad mountain basins, where the slope angle is likely to be greater than those of the Antarctic, but the length of the slope is much shorter. Within the United States, studies have been done in the Rocky Mountains (Banta 1982), in Washington at Rattlesnake Mountain (Horst and Doran 1982-3) and near Mt. Rainier (Buettner and Thyer 1966). Studies in Mexico (Fitzjarrald 1984) and in Europe (Defant 1949) have also been done. One of the typically smaller (i.e., occurring over a shorter slope length) atmospheric downslope flows that has been studied is along the surface of glaciers or snow patches (Ohata and Higuchi, 1979), where the length along the slope may only be a few hundred meters. The phenomena is known as glacier wind, and can develop over such short lengths because of the sharp slope angle and the sustained cooling provided by evaporation and the existence of ice. Glacier winds are often observed during summer months because of the advection of warm air over the

surface of the glacier from the surroundings creates sharp temperature gradients normal to the slope.

Atmospheric slope flows can exist on even smaller length scales than those mentioned above, but the likelihood of observing them occurring naturally is small. There are two reasons: (1) any ambient wind or turbulence will obscure the effect, (2) and the velocity and thickness of the flow layer are small. The dimensional analysis done in a later section of this paper provides the tools required to perform a convincing analysis of the conditions under which slope flows are going to be observed. For practical reasons laboratory scale experiments are done with fluids other than air and often involve concentration gradients and crystalization rather than temperature gradients for example (Turner and Ellison 1959, and Huppert et al. 1986). This paper is limited to atmospheric conditions.

Computational Studies

The computational advances made over the past 15 years have made obtaining numerical results from the complete set of governing equations an appealing alternative to seeking analytical solutions. Because of the strongly coupled, non-linear nature of the slope flow equations, analytical solutions may require restrictive assumptions. Several authors (Leslie and Smith 1974, Rao and Snodgrass 1981, and Arritt and Pielke 1987) have taken advantage of available computational capacity in investigating slope flows. In addition to the advantage of seeking solutions using the full governing equations, numerical solutions can incorporate boundary conditions that further confound any attempt to obtain a purely analytical solution to the same problem. For buoyantly driven flows, the vertical velocity and temperature field will most commonly be desired when ambient, synoptically driven or larger domain mesoscale winds are present. Analytical solutions are often simplified to avoid this necessary complexity. Computational techniques provide an avenue to investigate a broad class of conditions under which slope winds are observed.

Leslie and Smith begin with the general conservation of mass, momentum and energy equations, retaining only the terms pertaining to short mesoscale buoyancy driven flows where the Boussinesq approximation has been made. A stream function is introduced, and the resulting set of equations is integrated numerically. Turbulence is treated through a constant eddy diffusivity, K , equal to $4 \cdot 10^4 \text{ cm}^2 \cdot \text{s}^{-1}$. The particular two-dimensional geographic arrangement is a shallow flat sloping region trailing off into a horizontal region. The associated meteorology places a ceiling on the region of influence by the existence of a strong high level inversion. Simulated ground cooling provides the driving temperature gradient for a katabatic wind. The air is initially quiescent within the domain. The reported simulation results are consistent with general observations of buoy-

ancy driven winds, but are not quantitatively compared with observations.

Arritt and Pielke describe their simulation method in much the same way as Leslie and Smith. However, the intervening period of more than a decade has provided sufficiently faster computers which afford a more comprehensive description of turbulence. In this case a prognostic turbulent energy provides the parameterization on which the exchange coefficients (analogs of the eddy diffusivity mentioned above) are based. The geometry for which the simulation was verified is the geography of Rattlesnake Mountain. Unlike Leslie and Smith, the meteorology provides no high level inversion, only the low level inversion created by radiative cooling that provides the impetus for the nocturnal katabatic wind. A comparison of the vertical velocity and potential temperature obtained by the simulation to data reveals qualitative and reasonably good quantitative results.

The effects of the ambient wind on the boundary layer thickness and vertical wind profile is a separate issue addressed by Arritt and Pielke using their simulation model.

Analytical Studies

Two general approaches have been used to obtain analytical solutions for buoyantly driven winds: (1) hydrolic methods and (2) conservation equation methods. The hydrolic method provides information on the bulk properties of the buoyant layer. In contrast, the conservation equation techniques provide detailed information about the velocity and temperature profiles inside the layer. The hydrolic techniques are generalizations of open channel hydrolics which have added the buoyant term to the governing equations Defant (1951) and Ball (1956). Hydrolic methods account for the acceleration of the buoyant layer by entrainment through the interface with the ambient air. Some hydrolic models have relied on the assumption of a constant buoyant layer thickness, an assumption shown in this paper to be invalid for most common observations of slope wind. Manins and Sawford (1979) use the hydrolic formulation to provide a one-dimensional model for slope flow. Extensions of this technique have been used to study slope wind in the presence of ambient wind Fitzjarrald (1984). Kondo and Sato (1988) have reformulated the hydrolic method by adding the parameterization of the interfacial stresses between the ground and ambient air. This method treats the buoyant layer as a parcel rather than a hydrolic channel. The hydrolic technique is perhaps limited most by its inability to provide detailed information about the flow and temperature profile within the jet. The first analytical approach to use the conservation equations to study slope wind was presented by Prandtl (1942). Prandtl solved one of the limiting cases of slope wind, which is seldom observed, but one of the few cases that lends itself to a simple solution. Unfortunately, his simple solution to the model equations is the basis for comparison of much of the subsequent analysis in the field. Later analyses by Rao and Snodgrass (1980), Gutman (1983), as well as the computational work mentioned earlier were also developed directly from the conservation equations. Because detailed informa-

tion about the flow is often desired, the conservation method has received more attention in recent times.

Motivation for the Present Analysis

The analysis presented by the previous authors has served to describe many features of slope wind. However, the subject of slope wind itself is rather disjoint, lacking a comprehensive unification of the observations, theoretical developments, and computational results presented by previous authors. As a consequence, persons wishing to make quick estimates of the impact of buoyant winds at a particular location have to perform relatively detailed calculations or computer simulations for the case they are examining. Unfortunately, the comparisons made by many of the authors of their findings and those of previous authors was done for cases not in the same physical regime; thus, misleading conclusions that have been expressed about slope winds in general which should have only been made for specific flow regimes. For example, Rao and Snodgrass state without explanation, “The Coriolis effects are negligible except for very small slope angles.” The work of Mahrt (1982) attempts to tackle some of the problems by unifying the field, but falls short of this goal primarily because the work does not contain a comprehensive dimensional analysis. It only has scaling arguments. In particular, it side steps some of the fundamental issues of the dimensional analysis, such as determining the characteristic thickness for the buoyant air flow layer.

The comprehensive dimensional analysis done in this paper establishes the fundamental dimensionless groups necessary to perform a rapid preliminary analysis of slope wind. The potential impact of slope winds in a particular region can quickly be assessed based only on the fundamental characteristic parameters of slope length, slope angle, ambient temperature, atmospheric lapse rate, and surface to air temperature differential. Using these parameters, as indicated in this paper one can quickly establish such factors as (1) the characteristic depth of the buoyant layer, (2) the characteristic time frame required for flow to develop, (3)

the characteristic velocity, (4) the Reynolds number, (5) the importance of Coriolis forces through the Rossby number, (6) and the importance of entrainment through the bulk Richardson number. This boundary layer analysis provides a coherent picture of the interconnections between the model equations, computer simulations, and observations made under various conditions.

The ability to rapidly assess the impact of slope wind is important to the studies of other fields besides meteorology, e.g., air pollution and air quality. Individuals, in these and other fields, may wish to investigate the likelihood of slope wind occurring in a particular system without becoming an expert at the analytical and numerical methods for slope winds. The transport and dispersion resulting from slope winds in and surrounding populated areas is widespread, and the number of sites likely to be investigated for influence by buoyant wind effects is large. Finally, when more detailed studies are required, they should be undertaken with the appropriate equations for the particular regime of slope flow to be studied. The same holds true for observational field studies where *a priori* knowledge of the important parameters to measure will allow for the best characterization of the flow.

In the four sections that follow (1.2 - 1.5) an attempt is made to unify the subject of slope flow by performing a complete dimensional analysis. Sections 1.3 - 1.5 present the momentum, energy, and mass conservation equations. These equations are made dimensionless, noting which equations apply under which physical circumstances. Section 1.6 summarizes all of the dimensionless groups which were found.

1.2 Theory

Starting Equations

For momentum, the positionally averaged Navier-Stokes equation (1.2.1) which includes the effects of the earth's rotation and groups the Reynolds stresses into τ is the fundamental starting point for studying atmospheric buoyantly driven slope flow.

$$\rho \frac{D\mathbf{v}}{Dt} = -\nabla p - (\nabla \cdot \tau) + \rho g - 2\rho\Omega \times \mathbf{v} \quad (1.2.1)$$

The analogous equation for conservation of energy, balances the convective derivative of temperature with conduction and compressibility.

$$\rho C_p \frac{DT}{Dt} = -\nabla \cdot \mathbf{q} - p(\nabla \cdot \mathbf{v}) \quad (1.2.2)$$

As stated earlier, the density equations that motivate the flow are a function of temperature, thus 1.2.1 and 1.2.2 are coupled. The incompressible continuity equation (1.2.3) applies (Pielke, 1984).

$$\nabla \cdot \mathbf{v} = 0 \quad (1.2.3)$$

As a result, 1.2.2 can be simplified to eliminate the effect of local compression on the temperature field.

The shear stress tensor, τ , is composed of viscous, s , and turbulent, σ , (or Reynolds stress) components as denoted by Equation 1.2.4.

$$\tau = s + \sigma \quad (1.2.4)$$

As is most commonly done when modeling atmospheric transport, viscous transfer of momentum is neglected ($s \approx 0$) compared to turbulent transfer (Lumley and Tennekes, 1972). For the purpose of this analysis, the mixing length theory is applied in order to obtain an analytical representation of the remaining

turbulent term. The shortcomings of this approach are recognized, but as will be demonstrated below, the results obtained using this formulation are qualitatively representative of the observations. The mixing length hypothesis provides a simple way of representing turbulence using a constant exchange coefficient, K , as is shown in Equation 1.2.5.

$$\tau = -K\nabla v \quad (1.2.5)$$

By analogy, heat is transported according to the relationship given in Equation 1.2.6.

$$q = -K_T \nabla T \quad (1.2.6)$$

Equations 1.2.1 - 1.2.6 are a closed set of equations and provide the basis for studying buoyantly driven slope flow.

The Scaled Equations

As mentioned before, the absence of obvious characteristic parameters to use in scaling the governing equations may explain the lack of material in the literature on the appropriate way of approaching a dimensional analysis for slope flow. Two crucial areas lack obvious characteristic variables: lengths and velocities. While the slope length, L , provides a characteristic length along the slope, there is not a natural choice of a characteristic length vertical to, or when applicable, in the width dimension of the slope. The velocity has no natural choices in any of the three slope coordinates. With the characteristic length and velocity scales not explicitly available, one must attempt to construct them via other methods.

The work of previous authors provides insight into possible approaches to scaling the model equations. This includes the discussions presented by Prandtl, Mahrt, and the recent paper of Kondo and Sato, all mentioned previously. The

principal drawback of these presentations is that the full conservation equations were seldom the starting point for the analysis. In other words, terms in the model equations are disposed of or retained in an *ad hoc* fashion prior to actually doing a dimensional analysis. In Prandtl's presentation, terms appear to have been chosen partly on the basis of obtaining an easy analytical solution. Work since Prandtl has increasingly dealt with identifying which terms need to be retained to provide a realistic yet not analytically or computationally overburdening model for treating slope flow. Unfortunately, however, this decision process often preceeds doing any dimensional analysis. Consequently, the results are valid for the particular case that fits the starting equation, yet how this case fits into the bigger picture of slope flows in general is never made clear.

Many authors have treated slope flow in the way described above. The collective efforts which have been put forth, in a case by case way, provides some direction for doing a complete dimensional analysis. This previous work is put into context by this analysis.

In our method the characteristic scaling parameters are assumed to exist, and then obtained implicitly in terms of other known parameters through knowledge of the physical mechanisms of buoyantly driven slope flow. Having obtained the implicit characteristic scaling parameters in this way, it is then possible to completely specify dimensionless conservation equations for different regimes of buoyancy driven slope flow. This section, and the one following, serve to present the scaled equations and to define the characteristic constants, along with the pertinent dimensionless groups which arise.

Boundary Layer Analysis

The analyses presented below are derived in detail to elucidate the origin of the dimensionless groups and implicitly defined scaling parameters. The equations developed in Sections 1.3 - 1.5 build successively upon one another. In Section 1.3 the continuity equation (1.3.1) is scaled to show which forms are appropriate for long and short distances along the slope. In Section 1.4 the energy equation (1.4.1) is scaled and serves to implicitly define the vertical scaling thickness, Δ , for two regimes of the dimensionless number λ . In Section 1.5 the momentum equation (1.5.1) is non-dimensionalized. The complete set of dimensionless equations are developed for buoyantly driven slope flow in these sections.

1.3 Making the Continuity Equation Dimensionless

The continuity equation in the slope coordinates is made dimensionless first. Only one of the 6 characteristic parameters to do the scaling is known *a priori*, this is, L , the slope length. The unknown characteristic parameters are named, and subsequently defined in terms of the other parameters as physical constraints on the equations are enforced. When these parameters become known by correctly scaling the other conservation equations, they are then inserted into the definitions, and completely specifying the characteristic variables. The slope coordinate system is defined in Figure 1. The dimensional form of the equation is written out below.

$$\frac{\partial u}{\partial s} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial n} = 0 \quad (1.3.1)$$

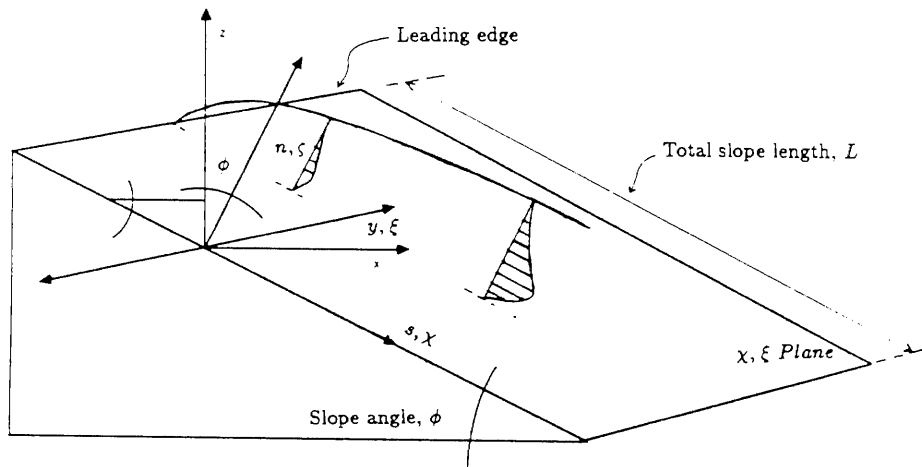


Figure 1. Slope coordinates relative to the rectangular coordinates of level ground.

The first step is to introduce the following, dimensionless variables:

$$\begin{aligned}
 U &= \frac{u}{U^*} \\
 \chi &= \frac{s}{L} \\
 V &= \frac{v}{V^*} \\
 \xi &= \frac{y}{L_y} \\
 W &= \frac{w}{W^*} \\
 \zeta &= \frac{n}{\Delta}
 \end{aligned} \tag{1.3.2 - 1.3.7}$$

Equation 1.3.1 is then rewritten in terms of the above, and divided through by $\frac{U^*}{L}$.

$$\frac{\partial U}{\partial \chi} + \frac{V^* L}{U^* L_y} \frac{\partial V}{\partial \xi} + \frac{W^* L}{U^* \Delta} \frac{\partial W}{\partial \zeta} = 0 \tag{1.3.8}$$

At this point, a little reasoning provides an assumption that saves a lot of trouble later and will be proven a little later in the discussion. Recall that at the leading edge of the boundary layer, the slope is uniform in the width dimension (by the original formulation of the problem), hence any fluid that is entrained into the boundary layer must be drawn downward from the air above the layer. Consequently, at the boundary layer's leading edge,

$$\frac{W^*}{\Delta} \gg \frac{V^*}{L_y} \tag{1.3.9}$$

and the appropriate form of the dimensionless continuity equation in this regime is

$$\frac{\partial U}{\partial \chi} + \frac{\partial W}{\partial \zeta} = 0. \tag{1.3.10}$$

The necessity that the remaining two terms balance forces the coefficient, $\frac{W^* L}{U^* L_y}$, to be equal to unity. This defines

$$W^* = \frac{\Delta}{L} U^* \tag{1.3.11}$$

in terms of the as yet undefined characteristic slope velocity and vertical scale length. The converse argument is possible when one looks at a slope flow that spans a long distance (long will be defined shortly). In this case,

$$\frac{W^*}{\Delta} \ll \frac{V^*}{L_y}, \quad (1.3.12)$$

the Coriolis forces, which would be hardly detectable over a short distance, have begun to bend the flow away from being colinear with the slope. Thus, over this distance the applicable continuity equation needs to be a balance of the gradients of the surface winds. In short this leads to the analogous relationships and equations to those presented above.

$$\frac{\partial U}{\partial \chi} + \frac{\partial V}{\partial \xi} = 0 \quad (1.3.13)$$

$$V^* = \frac{L_y}{L} U^* \quad (1.3.14)$$

For completeness it is also important to mention that when

$$\frac{W^*}{\Delta} = \frac{V^*}{L_y} \quad (1.3.15)$$

the full continuity equation,

$$\frac{\partial U}{\partial \chi} + \frac{\partial V}{\partial \xi} + \frac{\partial W}{\partial \zeta} = 0 \quad (1.3.16)$$

must be applied. In this case both 1.3.11 and 1.3.14 define the implicit characteristic velocities.

1.4 Making the Energy Equation Dimensionless

For buoyantly driven slope flows the energy equation represents a balance of slopewise advection with vertical transport of energy via turbulent diffusion. Though over long distances energy transport may be affected by the bending of the flow field due to Coriolis forces, the energy equation is primarily two-dimensional for the geometry being considered (a flat sloping surface with a near uniform surface temperature). It will be treated as such for all the model equations developed in this document. With this in mind

$$\rho C_p \frac{DT}{Dt} = -\nabla \cdot q - p(\nabla \cdot \mathbf{v}) \quad (1.4.1)$$

is combined with

$$q = -K_T \nabla T \quad (1.4.2)$$

and is written out in detail, eliminating the compressive term because of continuity, and keeping only the terms in the slope, s , and slope normal, n , directions as discussed above.

$$\rho C_p \left(\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial s} + v \frac{\partial T}{\partial n} \right) = K \left(\frac{\partial^2 T}{\partial s^2} + \frac{\partial^2 T}{\partial n^2} \right) \quad (1.4.3)$$

The relationship between the true temperature and the potential temperature is used, and the tranposition to the coordinates of the slope is made using Equations 1.4.4 - 1.4.6 below.

$$\frac{dT}{dz} = \frac{d\theta}{dz} - \Gamma \quad (1.4.4)$$

$$z = \chi \sin \phi \quad (1.4.5 \text{ and } 1.4.6)$$

$$z = \zeta \cos \phi$$

Additional dimensionless variables are introduced for the time and potential temperature.

$$\bar{t} = \frac{t}{t^*} \quad (1.4.7)$$

$$\Theta = \frac{\theta - \theta_\infty}{\theta_0 - \theta_\infty} \quad (1.4.8)$$

The dimensionless potential temperature, Θ , is defined in terms of the slope's surface potential temperature, θ_0 , and the upper air reference temperature, θ_∞ . Equation 1.4.3 is made dimensionless using the above, and those terms already defined in Equations 1.3.2 - 1.3.7, yielding,

$$\begin{aligned} \frac{1}{t^*} \frac{\partial \Theta}{\partial \bar{t}} + \frac{U^*}{L} U \frac{\partial \Theta}{\partial \chi} + \frac{\Gamma \sin \phi U^*}{\theta_0 - \theta_\infty} U + \frac{U^*}{L} V \frac{\partial \Theta}{\partial \zeta} - \frac{\Gamma \cos \phi U^*}{\theta_0 - \theta_\infty} \frac{\Delta}{L} V \\ = \frac{\alpha_T}{L^2} \frac{\partial^2 \Theta}{\partial \chi^2} + \frac{\alpha_T}{\Delta^2} \frac{\partial^2 \Theta}{\partial \zeta^2}. \end{aligned} \quad (1.4.9)$$

It is then rearranged, and terms of order $\frac{\Delta}{L}$ and smaller are lumped together as $O(\frac{\Delta}{L})$ terms.

$$\begin{aligned} \frac{L}{U^*} \frac{1}{t^*} \frac{\partial \Theta}{\partial \bar{t}} + U \frac{\partial \Theta}{\partial \chi} + V \frac{\partial \Theta}{\partial \zeta} + \frac{\Gamma \sin \phi L}{\theta_0 - \theta_\infty} U = \frac{\alpha_T L}{\Delta^2 U^*} \frac{\partial^2 \Theta}{\partial \zeta^2} \\ + O\left(\frac{\Delta}{L}\right) \text{ and smaller terms.} \end{aligned} \quad (1.4.10)$$

Defining the dimensionless group, $\frac{\Gamma \sin \phi L}{\theta_0 - \theta_\infty}$ as λ Equation 1.4.10 is rewritten.

$$\begin{aligned} \frac{L}{U^*} \frac{1}{t^*} \frac{\partial \Theta}{\partial \bar{t}} + U \frac{\partial \Theta}{\partial \chi} + V \frac{\partial \Theta}{\partial \zeta} + \lambda U = \frac{\alpha_T L}{\Delta^2 U^*} \frac{\partial^2 \Theta}{\partial \zeta^2} \\ + O\left(\frac{\Delta}{L}\right) \text{ and smaller terms.} \end{aligned} \quad (1.4.11)$$

When Equation 1.4.11 is written in this form it is apparent that the characteristic time for a slope flow to develop is $t^* = \frac{L}{U^*}$. Steady state is reached when t^* is small relative to the time frame of the flow duration, a condition that needs to be checked from case to case, but often holds for diurnal changes on the mesoscale.

At this point in the dimensional analysis a decision must be made as to which terms balance in order to define the different regimes under which various

forms of the dimensionless equation can be applied. For the slopeflow energy equation, there are two regimes which arise. To see how the two regimes arise, consider Equation 1.4.11.

The turbulent diffusion of energy to or from the ground's surface is an essential feature of the physical process, hence the first term right of the equality in 1.4.11 which embodies this effect is always retained. Excluding the unsteady term, the remaining terms that are left to balance the diffusive term are either of order 1 (the advective terms) or of order λ (the effect of transport through a stratified atmosphere). One can observe that λ scales with increasing length, L along the slope. Consequently, λ ranges from 0 at the slopes leading edge to some finite value depending on the length of the slope. For $\lambda < 1$ the advective terms balance the diffusive term. As a consequence, the coefficient of the diffusive terms must equal unity.

$$\frac{\alpha_T L}{U^* \Delta^2} = 1 \quad (1.4.12)$$

From this the vertical scaling parameter, Δ , is implicitly defined.

$$\Delta = \left[\frac{\alpha_T L}{U^*} \right]^{\frac{1}{2}} = L \left[\frac{\alpha_T}{\nu_T} \right]^{\frac{1}{2}} \left[\frac{\nu_T}{L U^*} \right]^{\frac{1}{2}} = L Pr^{-\frac{1}{2}} Re^{-\frac{1}{2}} \quad (1.4.13)$$

This leads to Equation 1.4.14, which is valid for $\lambda \leq 1$; this occurs at the leading edge of all buoyantly driven slopeflows.

$$\begin{aligned} \frac{\partial \Theta}{\partial \bar{t}} + U \frac{\partial \Theta}{\partial \chi} + \lambda U + V \frac{\partial \Theta}{\partial \zeta} - \lambda \frac{\Delta}{L} \cot \phi V &= \frac{\partial^2 \Theta}{\partial \zeta^2} \\ &+ O\left(\frac{\Delta}{L}\right) \text{ and smaller terms.} \end{aligned} \quad (1.4.14)$$

When $\lambda \geq 1$ the diffusive term must be of order λ .

$$\frac{\alpha_T L}{U^* \Delta^2} = \lambda \quad (1.4.15)$$

This results in the definition of the vertical scaling thickness for cases when $\lambda \geq 1$,

$$\Delta = Pr^{-\frac{1}{4}} \left[\frac{\alpha_T^2}{\Gamma(\sin^2 \phi) g \beta} \right]^{\frac{1}{4}} \quad (1.4.16)$$

and in the equation that is valid in this same regime.

$$\begin{aligned} \frac{1}{\lambda} \frac{\partial \Theta}{\partial t} + \frac{1}{\lambda} U \frac{\partial \Theta}{\partial \chi} + U + \frac{1}{\lambda} V \frac{\partial \Theta}{\partial \zeta} - \frac{\Delta}{L} \cot \phi V = \frac{\partial^2 \Theta}{\partial \zeta^2} \\ + O\left(\frac{\Delta}{L}\right) \text{ and smaller terms.} \end{aligned} \quad (1.4.17)$$

Equations 1.4.14 and 1.4.17 provide a set of equations that can be applied to slope flows over a full range of slope lengths. Equation 1.4.14 is always applied at the leading edge of the boundary layer regardless of whether λ exceeds 1. Equation 1.4.17 becomes applicable at the point along the slope when λ computed at that point is unity. This is analogous to the critical length, l_c , used by Kondo and Sato.

1.5 Making the Momentum Equation Dimensionless

The momentum boundary layer is a direct consequence of an unstably stratified density gradient above the surface of the slope. Thus the density term that embodies this effect in the momentum conservation equation must be balanced by the appropriate advective, diffusive and Coriolis terms in order to model the flow field correctly. This section presents the dimensional analysis that allows the correct balancing of terms to be obtained without the *ad hoc* speculation that has plagued some previous authors on the subject. A hierarchy of regimes will be found when this analysis is complete. Over short distances where Coriolis forces can be neglected, i.e., the Rossby number is large, two regimes exist as a result of restrictions imposed by the energy equation. These regimes are defined as regions along the slope where $\lambda \leq 1$ and $\lambda \geq 1$. Over long distances where Coriolis forces are important i.e., the Rossby number is small $Ro \leq O(1)$, there exists a third regime. Generally, λ is likely to be of $O(1)$ or larger in this regime, so it is not deemed necessary to subdivide small Ro like what was done above for large Ro .

The conservation of momentum equation,

$$\rho \frac{D\mathbf{v}}{Dt} = -\nabla p - (\nabla \cdot \boldsymbol{\tau}) + \rho \mathbf{g}, \quad (1.5.1)$$

can be separated into its horizontal component equations and similarly scaled as was done to the conservation equations of Sections 1.3 and 1.4. The hydrostatic approximation is made in the direction normal to the slope. The results are Equations 1.5.2 and 1.5.3.

$$\begin{aligned} \frac{\partial u}{\partial t} + u \frac{\partial u}{\partial s} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial n} = \nu_T \left[\frac{\partial^2 u}{\partial s^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial n^2} \right] + \\ - \frac{1}{\rho} \frac{\partial p}{\partial s} + g \frac{T - T_\infty}{T_\infty} \sin \phi v + 2\omega \sin \psi v - 2\omega \cos \psi w \end{aligned} \quad (1.5.2)$$

$$\begin{aligned} \frac{\partial v}{\partial t} + u \frac{\partial v}{\partial s} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial n} = \nu_T \left[\frac{\partial^2 v}{\partial s^2} + \frac{\partial^2 v}{\partial y^2} + \frac{\partial^2 v}{\partial n^2} \right] + \\ - \frac{1}{\rho} \frac{\partial p}{\partial y} + g \frac{T - T_\infty}{T_\infty} \cos \phi v + 2\omega \sin \psi u \end{aligned} \quad (1.5.3)$$

In addition, the relationship between the pressure gradient and the boundary layer's external flow, u_e and v_e is described by the following two relationships assuming the flow is inviscid outside of the buoyant layer.

$$-\frac{1}{\rho} \frac{\partial p}{\partial s} = u_e \frac{\partial u_e}{\partial s} \quad (1.5.4)$$

$$-\frac{1}{\rho} \frac{\partial p}{\partial y} = v_e \frac{\partial v_e}{\partial y} \quad (1.5.5)$$

The two dimensionless momentum equations applicable to the front edge of the boundary layer for $\lambda \leq 1$ and for $\lambda \geq 1$ are determined first. Recall that at the leading edge of the boundary layer the flow field is arguably two-dimensional in the s and n coordinates. Combining Equation 1.5.2 with 1.5.4, restricting the equations to two dimensions, non-dimensionalizing using the characteristic variables appropriate to the front end of the boundary layer (1.3.1 - 1.3.6, 1.3.10) and, lumping all terms of order $\frac{\Delta}{L}$ into $O(\frac{\Delta}{L})$ gives Equation 1.5.6.

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + W \frac{\partial U}{\partial \xi} = \tilde{U} \frac{\partial \tilde{U}}{\partial \chi} + Pr \frac{\nu_T L}{U^* \Delta^2} \frac{\partial^2 U}{\partial \chi^2} + \frac{g(\sin \phi) \beta (\theta_0 - \theta_\infty) L}{U^{*2}} \Theta + O\left(\frac{\Delta}{L}\right) \quad (1.5.6)$$

Additional information from the results of Section 1.4 can be used to simplify the coefficients of Equation 1.5.6. Ambient winds are now assumed to be quiescent. Notice that the coefficient of the diffusive term contains the dimensionless grouping 1.4.12 or 1.4.15, which is unity when $\lambda < 1$ and is λ when $\lambda \geq 1$. Also, since for turbulent flow the Prandtl number is approximately 1 the order of the diffusive coefficient depends only on λ . This results in Equations 1.5.7a ($\lambda \leq 1$) and 1.5.7b ($\lambda \geq 1$).

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + W \frac{\partial U}{\partial \xi} = Pr \frac{\partial^2 U}{\partial \chi^2} + \frac{g(\sin \phi) \beta (\theta_0 - \theta_\infty) L}{U^{*2}} \Theta + O\left(\frac{\Delta}{L}\right) \quad (1.5.7a)$$

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + W \frac{\partial U}{\partial \xi} = Pr \lambda \frac{\partial^2 U}{\partial \chi^2} + \frac{g(\sin \phi) \beta (\theta_0 - \theta_\infty) L}{U^{*2}} \Theta + O\left(\frac{\Delta}{L}\right) \quad (1.5.7b)$$

The characteristic velocity colinear with the slope is determined directly from Equations 1.5.7a and b. Since the buoyant term, $(\frac{g(\sin \phi) \beta (\theta_0 - \theta_\infty) L}{U^{*2}} \Theta)$, must be retained, it is balanced with the largest of the remaining terms. For Equation 1.5.7a, the largest remaining term has a lead coefficient of 1; consequently $\frac{g(\sin \phi) \beta (\theta_0 - \theta_\infty) L}{U^{*2}}$ is equated to 1. Solving for U^* defines the characteristic velocity,

$$U^* = [g(\sin \phi) \beta (\theta_0 - \theta_\infty) L]^{\frac{1}{2}} \quad (1.5.8)$$

Farther along the slope where λ exceeds 1, and 1.5.7b applies, the balance is between the buoyant force and the $O(\lambda)$ diffusive term. This leads to a modified definition of the characteristic velocity in this regime,

$$U^* = Pr^{-\frac{1}{2}} \left[\frac{g(\sin \phi) \beta |\theta_0 - \theta_\infty|^2}{\Gamma} \right]^{\frac{1}{2}}. \quad (1.5.9)$$

1.5.8 and 1.5.9 specifies the characteristic velocity U^* in terms of the measureable physical parameters of the particular slope and potential temperature gradient being examined for the two regimes of λ . Finally, the most simplified forms for the momentum equation for the two regimes are given as follows.

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + W \frac{\partial U}{\partial \xi} = Pr \frac{\partial^2 U}{\partial \chi^2} + \Theta + O\left(\frac{\Delta}{L}\right) \quad (1.5.10a)$$

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + W \frac{\partial U}{\partial \xi} = Pr \lambda \frac{\partial^2 U}{\partial \chi^2} + \lambda \Theta + O\left(\frac{\Delta}{L}\right) \quad (1.5.10b)$$

When Coriolis Forces Predominate

When a slope flow extends over lengths scales on the order of hundreds of kilometers, as is the case in Antarctica, the effects of Coriolis forces are evident in the horizontal bending of the flow from the coordinate of the slope. This

bending is accounted for in the dimensional analysis of the momentum equation along the slope through its coupling with the momentum equation in the cross slope direction. The continuity equation, which is appropriate on these longer slopes, 1.3.13, provides the relationship between characteristic velocities in the horizontal plane that allows a dimensional analysis to be done. Proceeding in a fashion similar to that done in the first part of this section, the model equations appropriate over long distances in a buoyantly driven boundary layer flow are developed.

Starting with 1.5.2 and 1.5.3, the dimensionless forms of these equations appropriate for slope flow influenced by Coriolis forces can be derived. To isolate the Coriolis effect, synoptic pressure gradients are dropped from the model equations. Using Equations 1.3.14, 1.3.2 - 1.3.7, and 1.5.8 or 1.5.9, the momentum equations are non-dimensionalized, and the smaller terms in the vertical direction eliminated.

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + V \frac{\partial U}{\partial \xi} = \frac{1}{Re} \left[\frac{\partial^2 U}{\partial \chi^2} + \frac{L^2}{L_y^2} \frac{\partial^2 U}{\partial \xi^2} \right] + \Theta + \frac{f L_y}{U^*} V. \quad (1.5.10)$$

$$\frac{\partial V}{\partial \bar{t}} + U \frac{\partial V}{\partial \chi} + V \frac{\partial V}{\partial \xi} = \frac{1}{Re} \left[\frac{\partial^2 V}{\partial \chi^2} + \frac{L^2}{L_y^2} \frac{\partial^2 V}{\partial \xi^2} \right] - \frac{f L}{V^*} U. \quad (1.5.11)$$

The only source of motion in Equation 1.5.11 arises from the Coriolis force, thus the advective terms must balance it. Consequently, using 1.5.13,

$$\frac{f L}{V^*} = \frac{f L}{U^*} \frac{L}{L_y} = \frac{1}{Ro} \frac{L}{L_y} = 1. \quad (1.5.12)$$

Solving for the characteristic length, L_y , in the cross slope direction, ξ , gives

$$L_y = \frac{L}{Ro}. \quad (1.5.13)$$

Using 1.5.13, and the knowledge that the Reynolds number, Re , is very large compared to 1 further reduces Equations 1.5.10 and 1.5.11 to those given below.

$$\frac{\partial U}{\partial \bar{t}} + U \frac{\partial U}{\partial \chi} + V \frac{\partial U}{\partial \xi} = \Theta + V. \quad (1.5.14)$$

$$\frac{\partial V}{\partial \bar{t}} + U \frac{\partial V}{\partial \chi} + V \frac{\partial V}{\partial \xi} = -U. \quad (1.5.15)$$

With this step completed the entire specification of all the pertinent dimensionless groups is done.

1.6 Characteristic and Dimensionless Slope Flow Parameters

The boundary layer analysis discussed above in Sections 1.3 - 1.5 gives rise to a set of dimensional scaling parameters and dimensionless groups. These can be used for assessing the impact of bouyantly driven flow in a particular geograpical area, from estimates or measurements of only a few important parameters: slope length, slope angle, ambient potential temperature, potential temperature difference, atmospheric lapse rate, and an atmospheric diffusivity. The parameters are denoted respectively as (L , ϕ , θ , θ' , Γ , and K). The dimensionless groups and scaling parameters are also useful in designing field studies and assessing the realism of laboratory scale experiments.

The characteristic, or scaling, parameters that arise from the analysis are discussed first. An attempt is made to introduce the implicit parameters in a logical, sequential order.

The energy equation serves to specify implicitly the characteristic boundary layer thickness, Δ , depending on the importance of other terms in the energy equation. As the buoyant wind drives the air to a different altitude, compressional heating or cooling can and will take place. The extent to which this will affect the buoyant flow depends completely on the air surrounding the jet, and the speed at which the jet rushes through the surrounding air. The dimensional analysis provides the dimensionless group, λ , which assesses the importance of the elevation change on the conservation of energy equation. The magnitude of λ determines which of the terms will be balanced.

$$\lambda = \frac{\Gamma \sin \phi L}{\Theta_0 - \Theta_\infty} = \frac{L}{l_c} \quad (1.6.1)$$

The critical length, l_c , is then defined explicitly.

Two definitions for the characteristic vertical length, Δ , exist. First when $\lambda \leq 1$, Equation 1.4.13 (in Section 1.4), the balance between advective and

diffusive mechanisms for energy transport yields,

$$\Delta = LPr^{-\frac{1}{2}}Re^{-\frac{1}{2}} \sim L^{\frac{1}{4}} \quad (1.6.2a)$$

where the Reynolds, Re , and Prandtl, Pr , numbers are defined in terms of the groups below. Since the Reynolds number scales as $L^{\frac{3}{2}}$ (when $\lambda \leq 1$), for buoyantly driven flows (see 1.6.7 below), one can quickly obtain the result that the characteristic boundary layer thickness, Δ , for the case $\lambda < 1$, grows as $L^{\frac{1}{4}}$. The second case is when $\lambda \geq 1$, and the balance between advective and diffusive mechanisms is replaced by a balance between the $O(\lambda)$ term and the largest diffusive term, Equation 1.4.15 (in Section 1.4). This leads to the definition of the characteristic thickness as

$$\Delta = Pr^{-\frac{1}{4}} \left[\frac{\alpha_T^2}{\Gamma(\sin^2\phi)g\beta} \right]^{\frac{1}{4}} \sim L^0. \quad (1.6.2b)$$

The L^0 dependancy of Δ indicates the effect of compressional heating or decompressional cooling eventually prevents the boundary layer thickness from growing along the slope. Note that this is the exact result obtained by Prandtl.

From the χ direction momentum equation comes the characteristic velocities for $\lambda \leq 1$ (1.6.2a) and for $\lambda \geq 1$ (1.6.2b),

$$U^* = Pr^{-\frac{1}{2}} [g(\sin\phi)\beta|\theta_0 - \theta_\infty|L]^{\frac{1}{2}} \sim L^{\frac{1}{2}}, \quad (1.6.3a)$$

$$U^* = Pr^{-\frac{1}{2}} \left[\frac{g\beta|\theta_0 - \theta_\infty|^2}{\Gamma} \right]^{\frac{1}{2}} \sim L^0, \quad (1.6.3b)$$

as a consequence of balancing the order 1, $O(1)$, advective and diffusive transport terms or the $O(\lambda)$ diffusive term with the buoyant term. Previous authors (Mahrt, Sato and Kondo) have recognized Equation 1.6.3a as being important to the study of buoyantly driven flows, but Equation 1.6.3b has not previously been identified with buoyantly driven flows. Equation 1.6.3b is suggested in the

original work of Prandtl, who notes that the speed becomes independent of slope length and length for these conditions. (Hereafter θ will refer to the quantity $|\theta_0 - \theta_\infty|$.) The characteristic velocity normal to the slope, where $\lambda \leq 1$, comes directly from the continuity equation and can be written in terms of the above definitions.

$$W^* = \frac{\Delta}{L} U^* = Pr^{-\frac{1}{2}} Re^{-\frac{1}{2}} U^* \quad (1.6.4)$$

When λ is bigger than 1 the velocity normal to the slope vanishes, because there is no entrainment.

To verify the correctness of U^* as the characteristic velocity, a comparison of U^* (using Equation 1.6.3a only) versus the observed maximum velocity, U_{Obs}^{Max} , is presented in Figure 1 for a number of observations (see Tables 1 and 2). The data is fit with a line obtained by linear regression,

$$U_{Obs}^{Max} = 0.625 U^*, \quad (1.6.5)$$

having the regression coefficient (Pearson r value) of 0.81. The regression is forced through the origin. Though this value of r is certainly indicative of a linear fit between U_{Obs}^{Max} and U^* , a better correlation might be obtained were more comprehensive data available. For instance, when vertical wind profiles are measured, the incremental vertical measurement heights are not guaranteed to capture the true maximum velocity. In addition, several of the papers from which the data were obtained did not provide comprehensive information on θ , $\Delta\theta$, ϕ , and L from which to compute U^* . In such cases estimates were made for the variables based on other information supplied, such as topographical maps, and the likely climate and ground conditions of the area where measurements were made. (Table 2 from Sato and Kondo supplied information on several locations in Japan, and additional information from Antarctica.) Hopefully, authors of the results from future investigations will adopt the convention of comprehensively reporting

Site no.	Site name	Location	Author
1	Scholl Canyon	USA	Shair
2	Mt. Tsurugi	Japan	[Ohata and Higuchi]
3	Rattlesnake Mountain, 7/1980	USA, Washington	Horst and Doran
4	Rattlesnake Mountain, 7/1981	USA, Washington	Horst and Doran
5	Hakatajima	Japan	[Imaoka]
6	Cobb Mountain	USA	[Dickerson and Gudiksen]
7	Glacier de St. Sorlin	France	[Martin]
8	South Park	USA, Colorado	Banta and Cotton
9	Glacier San Rafael	Chile	[Ohata et al.]
10	Mizuho Station	Antarctica	[Adachi and Kawaguchi]
11	Mizuho Station	Antarctica	Chiba and Kobayashi
12	Syowa Station	Antarctica	[Adachi]
13	Adelie Land	Antarctica	Gosink

Table 1. Locations, sites, and authors.

[] indicate data obtained directly from Kondo and Sato, Table 2.

Site no.	L (m)	ϕ (degrees)	θ (K)	θ (K)	U^* (ms ⁻¹)	U_{Obs}^{Max} (ms ⁻¹)
1	225	22	0.25	290	0.86	1.0
2	250	9	4.0-6.4	270	2.3-3.0	1.4-2.2
3	422	21	3.9	299	4.4	2.3
4	422	21	2.4	300	3.4	2.1
5	690	12.3	2.0	270	3.3	2.4
6	2400	10.6	5.0	270	8.95	1.2
7	2500	7	2.0-6.6	270	4.7-8.5	1.4-4.4
8	14000	3.3	4.5	320	10.3	4.0
9	40000	4.0	3.5	270	18.8	5.0
10	300000	0.15	7.2	260	14.6	14.0
11	300000	0.15	7.0	248	14.7	10.0
12	550000	0.31	3.8	260	20.6	15.4
13	800000	0.28	5.0	260	27.4	15.0-22.0

Table 2. Critical parameters, characteristic and observed velocities.

the necessary data for making the above correlation. Regardless of shortcomings in the data used for the correlation, the implicit definition of the characteristic velocity, U^* , does appear to be the correct natural scaling parameter for buoyancy driven flows. The correlated equation (1.6.5) is useful in approximating the maximum velocity along a slope with only a few parameters needing to be estimated or measured.

Knowledge of the characteristic velocity allows for a number of other parameters to be obtained. These include the characteristic time for the flow to develop, as well as dimensionless groups common to fluid mechanics in general.

The characteristic time for the transient portion of the flow (all boundary conditions being constant), $t^* = \frac{L}{U^*}$ comes from either the energy or the momentum equation and is defined in terms of U^* . Table 3 shows characteristic times obtained from the data used to generate Figure 2. The characteristic times span the range from minutes to nearly half a day. As was mentioned previously, most katabatic winds exist on the order of hours as governed by the diurnal cycle. Thus, winds which have characteristic times much smaller than several hours are generally in steady or pseudo steady state, and the dynamics which are observed are related to the dynamics of the boundary conditions. Examples are changes in ground and air temperatures due to radiation and air movement aloft of the katabatic boundary layer. The winds which have characteristic times on the order of hours are likely to be either developing or ceasing with the exception of one notable category. The katabatic winds of Antarctica, as Table 3 shows, require times exceeding a half day to develop. It must be kept in mind, however, that the proper conditions for strong katabatic winds to develop and reach steady state at the polar region may exist over several days. When this is not the case, the Antarctic katabatic wind will not reach steady state. Thus, the above definition of the characteristic time is useful in determining the dynamics

Katabatic Wind Speed

Observed Maximum versus Characteristic

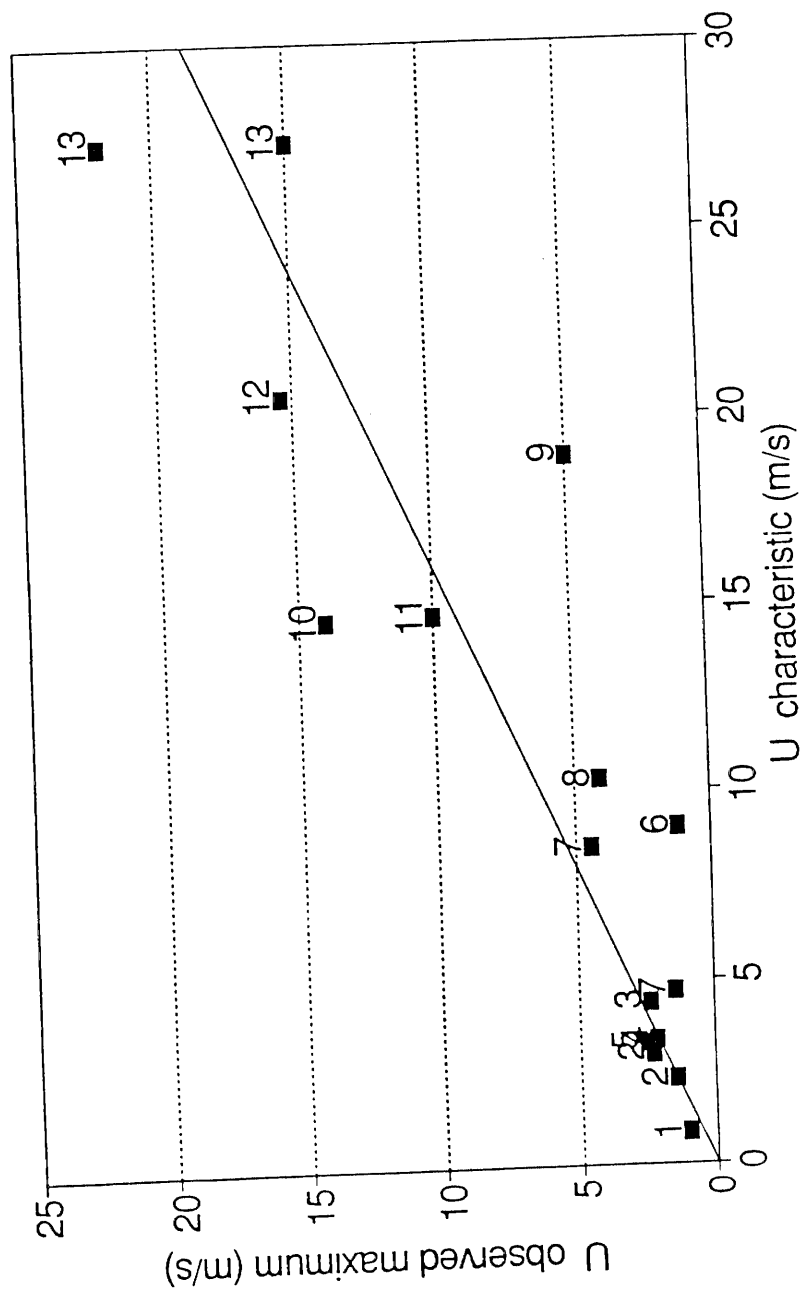


Figure 1. Linear correlation between the observed maximum velocity, U_{Obs}^{Max} , of the low level jet and the characteristic velocity, U^* , obtained via dimensional analysis. Numbers above the data points indicate the source of the data in Table 1.

Site no.	\bar{t} (minutes)	λ	Coriolis parameter $f, (\text{s}^{-1})$	Longitude $\psi, (\text{degrees})$	Rossby no. Ro
1	4.5	3.37	$1.3 \cdot 10^{-5}$	34 N	294
2	1.8	0.06-0.09	$1.3 \cdot 10^{-5}$	35 N	920
3	1.6	0.23	$1.7 \cdot 10^{-5}$	47 N	610
4	2.0	0.37	$1.7 \cdot 10^{-5}$	47 N	470
5	3.5	0.44	$1.3 \cdot 10^{-5}$	35 N	360
6	4.5	0.53	$1.6 \cdot 10^{-5}$	45 N	230
7	8.8	0.27-0.91	$1.8 \cdot 10^{-5}$	50 N	190
8	23	1.07	$1.5 \cdot 10^{-5}$	40 N	49
9	36	4.78	$-1.7 \cdot 10^{-5}$	47 S	28
10	342	0.65	$-2.2 \cdot 10^{-5}$	70 S	2.1
11	340	0.67	$-2.2 \cdot 10^{-5}$	70 S	2.1
12	445	4.69	$-2.2 \cdot 10^{-5}$	70 S	1.6
13	487	4.69	$-2.2 \cdot 10^{-5}$	70 S	0.81-1.2

Table 3. Characteristic time, λ , and Rossby number.

of katabatic winds, but should also be placed in the context of the dynamics of surrounding environment to avoid accidentally neglecting the time dependent terms of the conservation equations.

The familiar dimensionless groups to fluid analysis are summarized below, where the implicit definition of U^* is inserted as the characteristic velocity. Consequently, these dimensionless groupings take on some interesting forms and relations to one another. The Grashof number, Gr , is central to the study of bouyancy driven flows. Physically, the Grashof number represents a ratio of the buoyant forces to the viscous forces.

$$Gr = \frac{L^3 \rho^2 g (\sin \phi) \beta \theta}{K^2} \quad (1.6.6)$$

The Reynolds number, Re , which is the characteristic dimensionless ratio of inertial to viscous forces, can be written in terms of the Grashof number for buoyant slope flow as a consequence of the convective, viscous and buoyant forces all balancing in the analysis. Note that the turbulent kinematic viscosity is defined as $\nu_t = \frac{K}{\rho}$ and that two definitions are presented depending on the value of λ . The definition valid for $\lambda \leq 1$ is

$$Re = \frac{LU^*}{\nu_T} = Pr^{-\frac{1}{2}} \left[\frac{L^3 \rho^2 g (\sin \phi) \beta \theta}{K^2} \right]^{\frac{1}{2}} = Pr^{-\frac{1}{2}} Gr^{\frac{1}{2}} \sim L^{\frac{3}{2}}. \quad (1.6.7a)$$

The Reynolds number definition valid for $\lambda \geq 1$ is in terms of Equation 1.6.3b and l_c .

$$Re = \frac{l_c U^*}{\nu_T} = \frac{Pr^{-\frac{1}{2}}}{\nu_T} \left[\frac{g \beta \theta^4}{\sin^2 \phi \Gamma^3} \right]^{\frac{1}{2}} \sim L^0. \quad (1.6.7b)$$

It is not surprising that the Froude number, Fr , which represents the relative importance of the inertial forces to the gravitational forces, becomes independent of gravity as a result of the inertial forces being spawned by buoyant (i.e.,

gravitational) forces. For $\lambda \leq 1$

$$Fr = \frac{U^{\star 2}}{gL} = Pr^{-1}(\sin \phi)\beta\theta. \quad (1.6.8)$$

The Prandtl number,

$$Pr = \frac{\nu_T}{\alpha_T} = \frac{\rho C_p}{K_T} \frac{K}{\rho}, \quad (1.6.9)$$

which represents the ratio of the thermal diffusivity to the momentum diffusivity, is approximately 1 for turbulent flows, since the effective diffusive transport mechanism for both is the same. This may be used to simplify the previous dimensionless groups which include the Prandtl number.

The importance of Coriolis effects for a given geometry and differential temperature can be determined by computing the Rossby number, Ro . The Rossby number is specified in terms of the above characteristic variables, $\lambda \leq 1$,

$$Ro = \frac{U^{\star}}{fL} = \frac{l_{Co}}{L} = Pr^{-\frac{1}{2}} \left[\frac{g(\sin \phi)\beta\theta}{f^2 L} \right]^{\frac{1}{2}}, \quad (1.6.10a)$$

and for $\lambda \geq 1$,

$$Ro = \frac{U^{\star}}{fL} = \frac{l_{Co}}{L} = Pr^{-\frac{1}{2}} \left[\frac{g\beta\theta^2}{f^2 L^2 \Gamma} \right]^{\frac{1}{2}}, \quad (1.6.10b)$$

by the analysis performed in Section 1.5. Note the critical length along the slope at which Coriolis forces become important is called l_{Co} . The Coriolis parameter, f , is defined in terms of the earth's rotation rate, Ω , and the latitudinal angle, ψ , as $f = 2\Omega \sin \psi$. Table 3 contains the Rossby number for the observed slope flows. Note that $Ro \leq 1$ indicates Coriolis effects strongly affect the flow. For $1 \leq Ro \leq 20$ the Coriolis terms have a noticeable affect on the flow, while values of Ro greater than 100 indicate Coriolis effects are not likely to be observed in the particular circumstance being evaluated. Also, note that Ro is parenthetically

defined in terms of λ depending on the value of U^* which applies where Ro is being computed. Notice that the critical length, l_{Co} , after which Coriolis forces become dominant along χ (the slope) is defined in 1.6.10. The condition of l_{Co} being greater than L is equivalent to Ro being greater than 1. It may be clearer to use the critical length l_{Co} than to use $Ro \geq 1$ in some circumstances. For instance this allows direct comparison with the critical length l_c at which the effect of the buoyant jet traversing the stratified air becomes important for a given flow. Note that most slope flows have $Ro > 1$.

Once the Rossby number has been defined it is possible to explicitly define a characteristic velocity in the cross slope direction, V^* , which results from the Coriolis force. The cross slope scaling parameter width, L_y , comes from balancing terms in the ξ direction momentum equation, Section 1.5

$$L_y = \frac{L}{Ro} \quad (1.6.11)$$

$$V^* = \frac{L_y}{L} U^* = \frac{U^*}{Ro} = fL \quad (1.6.12)$$

Thus for a given location V^* is quickly determined in terms of the slope length and Coriolis parameter.

The last traditional dimensionless group is the Richardson number, Ri , and it can be expressed in terms of the above groupings by the following rearrangement.

$$Ri = -g(\cos \phi)\beta\theta \frac{\Delta}{U^{*2}} \quad (1.6.13)$$

The Richardson number is a measure of the stability of a stratified shear flow. The Richardson number presented above is the bulk form.

When thermally driven slope winds coexist with ambient winds, as they often do, it is important to ascertain the influence that the interaction has on transport. Two principle interactions summarize the extremes, the ambient wind can flow

counter current ,i.e., oppose the slope flow, or it can flow cocurrent ,i.e., with the slope flow. Obviously the effects of cross winds require a three-dimensional formulation, but such cases are not treated herein. The dimensionless group governing this interaction is a natural choice, \ddot{U} , in that it is a ratio of the ambient flow velocity to the characteristic velocity of the slope flow, for $\lambda \leq 1$

$$\ddot{U} = \frac{U_{\infty}}{U^*} = \left[\frac{U_{\infty}^2}{g(\sin \phi)\beta\theta L} \right]^{\frac{1}{2}}. \quad (1.6.14)$$

A similar relationship can be developed using the characteristic velocity for $\lambda \geq 1$ using Equation 1.6.3b. The magnitude of \ddot{U} indicates whether the buoyantly driven wind is a first or second order effect relative to the ambient wind. When $|\ddot{U}| < 1$ the buoyantly driven wind is the first order effect and the low level jet is likely to be observed. When $|\ddot{U}| > 1$ the ambient winds dictate the nature of the flow field. In this case buoyantly driven winds perturb the ambient flow field and their effect may be obscured.

Conclusions

Probably the most striking feature of buoyantly driven slope flows brought out by this analysis is the functional relationship between the boundary layer thickness and the slope length (1.6.2a and 1.6.2b). The two regimes are controlled by the mechanisms by which energy flows across the buoyant layer. At the leading edge of the slope, advective and diffusive transport balance, while further along the slope diffusive transport is balanced by the heating (or cooling) effect of entraining ambient air into the momentum boundary layer. The demarcation point between the two regimes is set by the value of the parameter λ . For $\lambda \leq 1$ the former case above holds, while for $\lambda \geq 1$ the latter case applies. Of course, in the overlap region around $\lambda = 1$ all three mechanisms for energy transport are in balance. It is very important when assessing the potential for slope flow to first calculate λ . The steeper and shorter the slope or the smaller the ground to air temperature gradient, the more likely λ will exceed 1. Based on λ 's value, one can calculate the remaining parameters of interest.

The growth of the buoyantly driven boundary layer at its leading edge as a function of the slope length to the quarter power ($\Delta \sim L^{\frac{1}{4}}$) is one aspect that has important implications with regard to pollution transport. The reason for this lies in the connection between boundary layer thickness and the entrainment of air from above the layer. A boundary layer which grows as $\chi^{\frac{1}{4}}$, where χ is used instead of L since one can choose to scale the slope length anywhere along the slope, entrains a large fraction of air at first and less and less with increasing distance along χ . This means that a buoyant boundary layer acts like a pump, preferentially extracting air at its leading edge and depositing it at the end of the slope.

This has two major effects on the transport of pollutants, depending on their origin relative to the buoyant flow. First, if elevated pollution levels exist at the

front of the buoyant boundary layer, the polluted air is drawn in and transported within the layer until the end of the slope is reached. Dilution of the pollutant by air from above the layer farther along the slope is limited because the slow growth of the layer imposed by the $\chi^{\frac{1}{4}}$ law provides minimal entrainment into the layer. If the slope is long enough such that λ exceeds 1 ($\Delta \sim L^0$), then the boundary layer thickness becomes constant, i.e., no entrainment into the layer from above, and dilution is limited completely to turbulent transport across the boundary layer. Second, if the slope along which the boundary layer develops is a source of pollutant, for instance a landfill, then by the same arguments presented above, the pollutant is trapped and transported within the boundary layer. In this way the emissions from a ground source can be moved at ground level from one location with limited dilution to another location.

The dimensional analysis can be used to determine the kind of interaction the slope wind will have with the ambient wind. The nature of the slope wind-ambient wind interaction has been discussed in several papers. The observations presented therein substantiate the persistence of slope wind in the presence of ambient wind. Interpretation, and use of equation 1.6.14, is by an attempt to provide a quick method for estimating whether a slope wind will persist and be observable in the presence of cocurrent or counter current ambient wind.

The following guidelines are proposed when interpreting the value of \tilde{U} . For $|\tilde{U}| \gg 1$ the effects of the slope wind will not be observed. Under such circumstances the atmospheric conditions that produce slope wind are not likely to be maintained. When $|\tilde{U}| \approx 1$ the effect of the slope wind may be evident as a distortion in the logarithmic ground level wind velocity profile, but a true jet will not be observed. When $1 > |\tilde{U}| > 0$ the observed wind profile will have a low level jet, but will also show the influence of the ambient wind with increased vertical height. Naturally, when $\tilde{U} = 0$ ambient winds are not present and the

low level jet will be observed undisturbed. Because observations available from the literature are insufficient to substantiate these criteria, the above are simply guidelines and may need to be revised as more data becomes available.

Slope winds are often nested within other mesoscale meteorological regimes, for example, the mountain-valley system. Thus it may be possible to obtain estimates of the ambient wind potential in an area also susceptible to slope wind, by techniques developed for these other mesoscale regimes. This information, plus the pertinent parameters available for the slope wind can be combined and used in Equation 1.6.14 to serve as a predictive tool for estimating the near ground behavior of winds along a slope.

It is useful to place several previous papers into the context of the dimensional analysis just presented. Examination of Prandtl's solution reveals that it is the solution of a slope flow for the dimensionless case of $\lambda \gg 1$. In other words, Prandtl's solution is a realistic solution of the extreme case where entrainment of the ambient air leads to a reversal of the potential temperature gradient atop of the boundary layer. If Ohata and Higuchi (1979) had known this, they probably would not have expected their observations on a snow patch, which had a λ of 0.04 – 0.06 (Table 3), to match well with Prandtl's solution. The paper by Rao which suggests that Coriolis effects are negligible except for small slope angles should be amended based on Equation 1.6.10. The statement should also include cases where the slope is long, the ground to ambient air temperature gradient is small, or the Coriolis parameter is large; specifically, cases where the Rossby number is 1 or smaller.

Acknowledgements

The assistance of M. Deem and P. Wyckoff in the preparation of this document was greatly appreciated.

1.7 Application of the von Karman – Pohlhausen Method to Bouyantly Driven Boundary Layers

Introduction

The boundary layer equations developed in Sections 1.2 - 1.5 cannot, in general, be solved to give closed form solutions except for the case when the critical slope parameter, λ , is much larger than one (Prandtl, 1952). Atmospheric slope flows exist for smaller values of λ , and so it would be desirable to have a method of acquiring the detailed temperature and velocity profiles for these conditions. Computer generated profiles are possible but not always practical, and hydrolic techniques do not allow for the computation of temperature and velocity profiles within the slope flow boundary layer.

The boundary layer method devised by von Karman and Pohlhausen, and described by Schlichting (1979), is an alternative approximate solution strategy capable of satisfying the above mentioned criteria. Through the method the detailed temperature and velocity profiles are obtained, often without the need for any computer assistance. The drawback of the technique is that it is approximate. However, the error introduced in other similar problems is generally small, usually less than ten percent, when implemented as it will be below. Some of the physical parameters on which any solution would rely are not known more accurately than to ten percent, and field measurements of the slope wind speed profiles contain errors of this magnitude. The ability to obtain the velocity and temperature profiles far outweighs the small error introduced by this method when it is utilized for comparison with field measurements and modeling atmospheric conditions.

Method

The boundary layer equation for energy (Equation 1.4.11), and for momentum (Equation 1.5.6), are integrated from the ground, 0, to the edge of the slope flow boundary layer at $\delta(\chi)$. A uniform speed external to the flow, U_∞ , is assumed. It has been made dimensionless by the characteristic slope velocity, U^* . The resulting equations are combined with the continuity equation, and Liebniz's differentiation of an integral rule is applied. After rearrangement this gives Equation 1.7.1 and 1.7.2. These equations are very similar to the standard boundary layer integral equations for flow over a flat plate, with the addition of the terms which couple the momentum to the energy.

Energy

$$\frac{d}{d\chi} \int_0^{\delta_T(\chi)} U \Theta \, d\zeta + \lambda \int_0^{\delta_T(\chi)} U \, d\zeta = - \frac{\partial \Theta}{\partial \zeta} \Big|_{\zeta=0} \quad (1.7.1)$$

Momentum

$$\frac{d}{d\chi} \int_0^{\delta(\chi)} U(U - U_\infty) \, d\zeta = -Pr \frac{\partial U}{\partial \zeta} \Big|_{\zeta=0} + \int_0^{\delta(\chi)} \Theta \, d\zeta \quad (1.7.2)$$

The δ 's are made dimensionless in order to be compatible with the dimensionless forms of the boundary layer equations and are defined below.

$$\delta_T(\chi) = \frac{\delta_{Therm}}{\Delta} \quad (1.7.3)$$

$$\delta(\chi) = \frac{\delta_{\rho v}}{\Delta} \quad (1.7.4)$$

The boundary layer thicknesses, δ_{Therm} for the thermal layer, and $\delta_{\rho v}$ for the momentum layer, are dimensional. To complete the specification of the method, polynomial profiles are assumed for the temperature and velocity and fit to the boundary conditions which are known or may be imposed by physical arguments.

Boundary Conditions and Profiles

I. Temperature: Define $\eta \equiv \frac{\xi}{\delta(x)}$, $\eta_T \equiv \frac{\xi}{\delta_T(x)}$, and $h \equiv \frac{\delta_T(x)}{\delta(x)}$

Boundary Conditions:

1. $\Theta(\eta_T = 0) = 1$
2. $\Theta(\eta_T = 1) = 0$
3. $\Theta'(\eta_T = 1) = 0$
4. $\Theta''(\eta_T = 0) = 0$

Meaning of the Boundary Conditions

See Section 1.4 Equation 1.4.8 which defines the temperature variable, Θ . Boundary condition 1 and 2 specify the dimensionless potential temperature difference function at ground level and at the edge of the thermal boundary layer. Boundary condition 3 guarantees the temperature profile joins the external temperature field smoothly. Boundary condition 4 is obtained directly from Equation 1.4.14.

Assumed profile:

$$\Theta = a_1 + a_2\eta_T + a_3\eta_T^2 + a_4\eta_T^3$$

After applying the boundary conditions:

$$\Theta = 1 - \frac{3}{2}\frac{\eta}{h} + \frac{1}{2}\left(\frac{\eta}{h}\right)^3 \quad (1.7.5)$$

II. Velocity (quiescent external flow): Define $\eta \equiv \frac{\xi}{\delta(x)}$

Boundary Conditions:

1. $U(\eta = 0) = 0$
2. $U(\eta = 1) = 0$
3. $U'(\eta = 1) = 0$
4. $U''(\eta = 0) = -Pr^{-1}$

Meaning of the Boundary Conditions

See Section 1.5 Equations 1.5.8 which defines the dimensionless velocity, U . Boundary condition 1 and 2 specify the dimensionless velocity at ground level and at the edge of the momentum boundary layer is zero. Boundary condition 3 guarantees the velocity profile joins the external velocity field smoothly. Boundary condition 4 is obtained directly from Equation 1.5.10a.

Assumed profile: $U = b_1 + b_2\eta + b_3\eta^2 + b_4\eta^3$

After applying the boundary conditions:

$$U = \frac{\delta^2}{4Pr} \eta(1 - \eta)^2 \quad (1.7.6)$$

III. Velocity (Uniform external flow): Define $\eta \equiv \frac{\zeta}{\delta(x)}$

Boundary Conditions:

1. $U(\eta = 0) = 0$
2. $U(\eta = 1) = U_\infty$
3. $U'(\eta = 1) = 0$
4. $U''(\eta = 0) = -Pr^{-1}$

Meaning of the Boundary Conditions

Same as the velocity conditions in II. above, except boundary condition 2 specifies the dimensionless velocity at the edge of the momentum boundary layer is U_∞ .

Assumed profile: $U = b_1 + b_2\eta + b_3\eta^2 + b_4\eta^3$

After applying the boundary conditions:

$$U = \left[\frac{3}{2}U_\infty + \frac{\delta^2}{4Pr} \right] \eta - \frac{\delta^2}{2Pr} \eta^2 - \left[\frac{1}{2}U_\infty - \frac{\delta^2}{4Pr} \right] \eta^3 \quad (1.7.7)$$

Velocity and Temperature Boundary Layer Thickness

Inserting the assumed profiles, i.e., Equations 1.7.5 and 1.7.6, into the integral form of the boundary layer equations and evaluating the integrals leads to a set of equations for the unknown functions, $\delta(\chi)$, $\delta_T(\chi)$, or $h(\chi)$. The solution of this set of equations requires additional assumptions since there are 3 unknown functions and only 2 equations. The assumption that the ratio of the boundary layer, h , is a constant, at least to first order, can be argued on the basis that the existence of the momentum boundary layer results directly from the thermal boundary layer. This leaves the following pair of differential equations to be solved and manipulated. The Prandtl number is assumed to be unity.

From the energy equation comes Equation 1.7.8.

$$\delta(\chi)^3 \frac{d\delta(\chi)}{dx} = \frac{Pr}{2} \left(\frac{h^3}{40} - \frac{h^4}{48} + \frac{3h^5}{560} \right)^{-1}. \quad (1.7.8)$$

From the momentum equation and velocity profile, Equation 1.7.6, comes Equation 1.7.9.

$$\delta(\chi)^3 \frac{d\delta(\chi)}{dx} = 336Pr^2 \left(\frac{3}{4} - \frac{3}{4h} + \frac{1}{8h^3} \right) \quad (1.7.9)$$

The boundary condition is $\delta(\chi = 0) = 0$.

From the momentum equation and velocity profile, Equation 1.7.7, comes Equation 1.7.10.

$$\frac{d\delta(\chi)}{dx} = \frac{-\frac{3}{2}PrU_\infty + \delta^2 \left(\frac{3}{4} - \frac{3}{4h} + \frac{1}{8h^3} \right)}{a_1 U_\infty^2 \delta + a_2 U_\infty \delta^3 + Pr^{-2} \delta^5} \quad (1.7.10)$$

The constants are defined as: $a_1 = 0.6107$, $a_2 = 0.05895$, $a_3 = 0.002976$ and the boundary condition is $\delta(\chi = 0) = 0$. This relationship can be used to study the influence of an external wind on the boundary layer thickness.

The value of h is the first piece of information available directly. By dividing Equation 1.7.8 by Equation 1.7.9, rearranging, and then solving the resulting

fifth order polynomial for its zeros, a set of possible solutions for h are obtained. The most probable value is $h = 1.05$. The other possible values to choose from were either complex or physically unrealistic. With h known, $\delta(\chi)$ can now be evaluated from Equation 1.7.9 and the boundary condition for δ . This gives:

$$\delta(\chi) = 3.7226\chi^{\frac{1}{4}}. \quad (1.7.11)$$

Maximum and Average Velocity

The average velocity is found from the approximate velocity profile, Equation 1.7.6, by integration.

$$U^{Ave}(\chi) = \int_0^1 U(\eta) d\eta = \frac{\delta^2(\chi)}{48Pr} \quad (1.7.12)$$

The maximum velocity is found from the approximate velocity profile by differentiation. It occurs at $\eta = \frac{1}{3}$, i.e., one third the total boundary layer thickness.

$$U^{Max}(\chi) = \frac{\delta^2(\chi)}{27Pr} \quad (1.7.13)$$

The ratio of the maximum velocity to the average velocity is

$$\frac{U^{Max}}{U^{Ave}} = \frac{48}{27} = 1.77. \quad (1.7.14)$$

Dimensional Relationship of Velocity Maximum

The maximum dimensional velocity, U_{Dim}^{Max} , is related to the dimensional boundary layer thickness as:

$$U_{Dim}^{Max}(s) = \frac{U^{\star 2}}{27LK} \delta_{\rho v}^2(s). \quad (1.7.15)$$

The dimensional slope coordinate is s . This is found by taking Equation 1.7.13 and using the relationships found in Section 1.5 to reconstitute a dimensional equation. This relationship holds for $\lambda \leq 1$. The parameter K is the effective eddy diffusivity for momentum.

Comparison of the Velocity Profile to Field Observations

During field measurements sufficient data has not been collected to do a comprehensive comparison between observations and the boundary layer analysis previously described. Though numerous field investigations of slope winds have been done, none of them have thoroughly studied the development of the buoyant boundary layer along the slope. Most downslope studies have been at the end of the slope, where the winds are the strongest, and thus at the end of the boundary layer. From an observational point of view, this is where the speed is at a maximum, and so it makes sense to take measurements there. From a modeling and verification-of-theory point of view, measurements should be taken at numerous places along the slope.

The assumption that slope flows can be approximated by a nearly constant thickness all along the slope must have been used to design most field studies. This assumption appears never to have been tested except in the large scale slope flows in Antarctica. The boundary layer analysis given in Sections 1.2-1.5 describes the different regimes for slope flow and shows this approximation is only valid for $\lambda \gg 1$. Whereas this assumption is valid in the flows in Antarctica, it does not hold for many of the studies which have been done on smaller scales.

A consequence of assuming the boundary layer thickness is constant is the ability to neglect the evolution of the flow along the slope when taking measurements. Remember from Section 1.5 and 1.6 that the boundary layer and slope wind profile have become independent of the slope coordinate. This assumption is not valid case for all regimes of slope wind. However, some investigators have treated measurements as if they are in this regime irrespective of whether they are or not. In no observational paper read to date has the author attempted to describe the slope flow regime on the basis of the fundamental characteristic scalings. Not surprising is fact that the measurements are then being compared

to models of the wrong physical regime of the slope flow. Because of this, the conclusions drawn by some authors about the applicability of their models are still to be proven.

Unfortunately, the model proposed above cannot be thoroughly tested either because of the absence of measurements at multiple places along a single slope. Only a fit of theory and profiles are possible at single locations along the slope. One plot is shown here, Figure 1, where the model was compared to measurements reported by Banta and Cotton (1982). The plot is generated by inserting the measured boundary layer thickness in Equation 1.7.6, using a slope length of 3300 meters, slope angle of 3.3 degrees, and a potential temperature difference of 4 degrees. This data could be extracted from plots in Banta and Cotton's paper. The model produces results which are qualitatively correct, but to draw a comparison between this model and other models would be inappropriate. When better data exists such comparisons can be undertaken.

One point on which it is appropriate to draw a conclusion is the correctness of assuming the maximum boundary layer speed will occur one third the height up from the ground. From the plot this seems to be a good assumption. This has also been noted in other comparisons not reported herein.

Comparison of Boundary Layer Profiles

Measurements & Karman-Pohlhausen Method

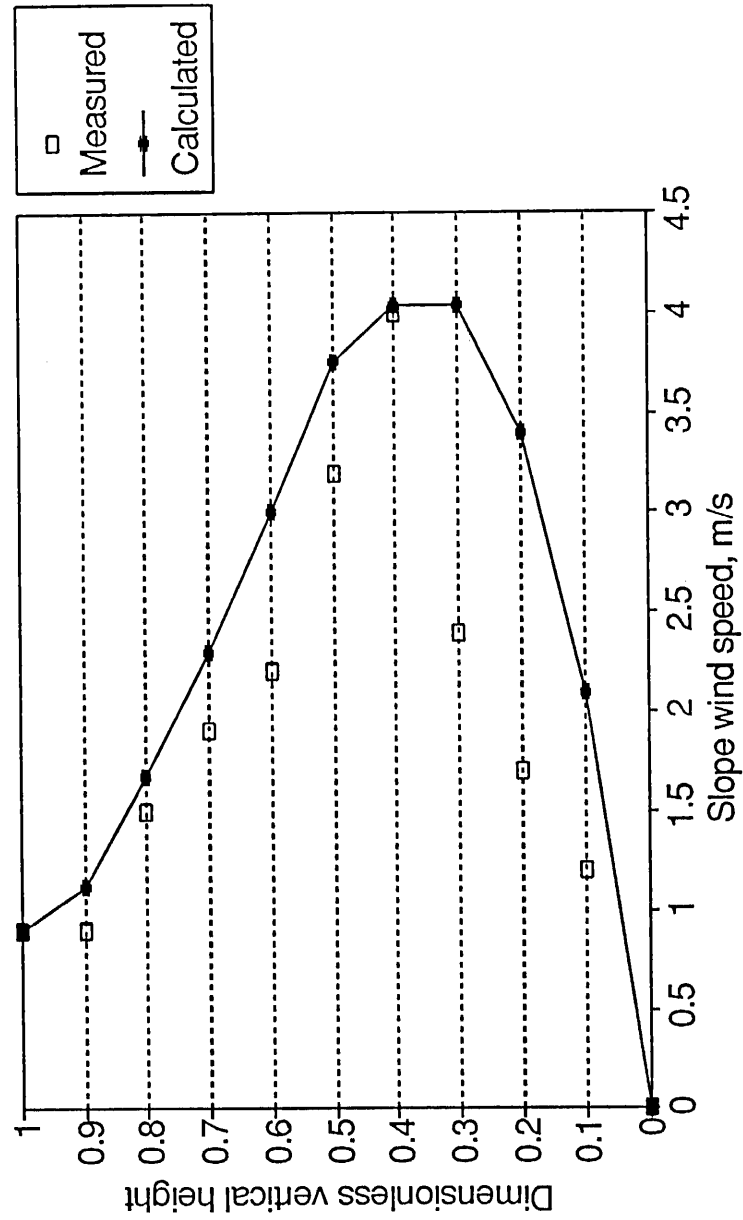


Figure 1. Comparison of the velocity profile given by Equation 1.7.7 with measurements made by Banta and Cotton (1982).

Effect of an External Wind on the Boundary Layer Thickness

The flux of momentum towards the ground resulting from an external wind has the capacity to change the thickness of the boundary layer along the slope. Equations 1.7.7 and 1.7.10 are used to study the quantitative changes in the boundary layer thickness resulting from a uniform external flow. The influence is studied for two reasons: (1) most often slope winds occur in the presence of a breeze or wind, and (2) changes in the boundary layer thickness affect the dilution of emissions in the layer.

To study the change in boundary layer thickness, Equation 1.7.10 is integrated numerically to the end of the slope, i.e., the boundary layer. For the dimensionless equation this is from $\chi = 0$ to 1. All boundary layer thicknesses reported below are given at $\chi = 1$. Seven simulations were done for external wind speeds which ranged from -0.5 (minus indicates opposing flow) to 0.2 of the characteristic slope velocity. Figure 2 shows the results of these simulations plotted as the boundary layer thickness δ normalized by the boundary layer height when $U_\infty = 0$ (Equation 1.7.11). The simulations were done using a dynamic simulation package, ACSL, using a fourth order Runge-Kutta as the numerical integration scheme. One assumption which was implicit in the simulation was the value of the ratio of the thermal to momentum boundary layer thickness, h , was constant. In the presence of an external wind, the momentum boundary layer thickness is no longer completely determined by the thermal layer. No method was able to be devised to determine how the error in this assumption grew with increasing U_∞ . However, this assumption is valid in the neighborhood of the external wind speed being zero, $U_\infty = 0$. It was extended to the range of U_∞ used in the calculations.

In the following discussion when the boundary layer is referred to, this means the end of the boundary, $\chi = 1$. Since h was found to be approximately equal to

External Flow Effects on Boundary Layer Thickness for Cocurrent & Opposing Flow

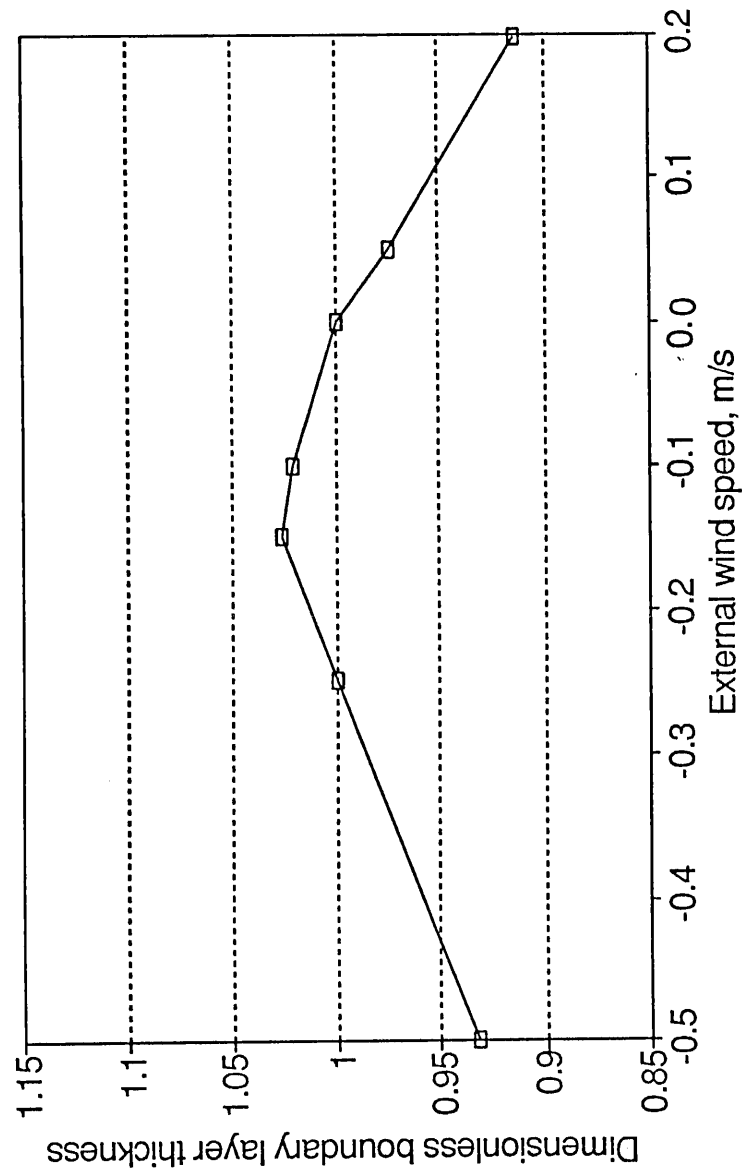


Figure 2. Changes in the slope flow momentum boundary layer thickness in the presence of an external constant flow. Both opposing flows and flows moving with the boundary layer are depicted. Notice the maximum height occurs for a small opposing flow. The external wind speed is made dimensionless by the characteristic slope flow speed, U^* .

one, the thermal and the momentum boundary layer are interchangeable. The phenomena being discussed are for the entire boundary layer, but the calculations are examined at the end of the layer.

A maximum in the boundary layer thickness occurs for an opposing flow which is flowing fifteen percent of the characteristic slope flow speed, U^* . Ostensibly two parameters control the thickness of the layer: (1) the height of the velocity maximum above ground and (2) the magnitude of the maximum velocity. Knowing that a maximum boundary layer thickness occurs, the interaction of the two factors can be explained as follows. For small opposing flows little momentum is being diffused into the layer to reduce the maximum velocity. Under these conditions it takes the original length $\delta(U_\infty = 0)$ to reach zero speed at the outer edge of the boundary layer, plus a little amount more to go from a speed of zero to the small opposing speed. At increased speeds the opposing flow sends more momentum into the layer. This retards the speed at the center of the flow, thus the maximum velocity is smaller and lower to the ground. These two effects are in balance at the maximum, and then larger opposing flow causes the boundary layer to decrease in size. The boundary layer has the same thickness when $U_\infty = 0.25U^*$ as it does when there is no external flow.

When the external flow is in the same direction as the slope flow, the boundary layer thickness decreases. In this case the flux of momentum into the top of boundary layer allows it to reach the external flow speed across a shorter distance. When the external speed roughly reaches the slope speed, the boundary layer will again need to grow because the transfer of momentum from the external flow will begin to be slowed by momentum transfer to the ground.

In either case, i.e., opposing or similar flows, when the external flow speed is approximately greater than the characteristic slope speed, the boundary layer thickness will be determined by the opposing flow.

Conclusions

Five major conclusions can be drawn from the previous section.

1. The von Karman-Pohlhausen technique is successfully applied to slope flows.
2. The velocity profile which is obtained compares well with measurements.
3. The slope flow boundary layer thickness grows slightly in a breeze opposing the direction of its flow, reaching a maximum height when the external flow is fifteen percent the size of the characteristic slope flow speed, $U_\infty = 0.15U^*$.
4. The slope flow boundary layer thickness decreases steadily in a breeze flowing in the same direction as the flow.
5. A carefully designed, comprehensive slope flow field study will have to be done to determine the adequacy of the above technique. This field study should consider the following advice:
 - a. Use the results from the dimensional analysis to assess/pick a site for the measurements. A site with a gradual slope on the order of 10 km would be ideal, perhaps a glacier.
 - b. Do simultaneous measurements at many locations along the slope. Make certain the leading edge of the flow can be located by first doing diagnostic measurements.
 - c. Estimate the boundary layer thickness for the slope flow *a priori*, confirm it with diagnostic measurements, and then do extensive measurements from ground level up to this height including a reasonable margin for variations. Definitely find the maximum speed at each point along the slope, since testing the theory depends on knowing it.
 - d. A simultaneous atmospheric tracer release might be considered to parameterize turbulence, mixing, and transport in and of the flow. If real time visualization of the slope flow were being done by reconstructing the data as it were being taken, a very controlled tracer release into the

leading edge of the flow would be possible.

1.8 Modeling the Evolution of a Two – Dimensional Thermally Driven Atmospheric Boundary Layer

Introduction

The objective of this section is to present a non-numerical approach for modeling the transient development of a buoyantly driven slope flow boundary layer. No prior work in this area exists in the literature except for that done by computer simulations, e.g., Leslie and Smith (1974), Arritt and Pielke (1986). The approach to be considered is to extend the von Karman-Pohlhausen method discussed in Section 1.7 to incorporate the transient terms of the momentum equation. A similar approach was used to analyze the impulsive motion of a semi-infinite flat plate at speed U , by Stewartson (1950). In this study, Stewartson determined the boundary layer development beyond an essential singularity at $Ut = x$, away from the leading edge, was independent of the horizontal coordinate, x . By analogy slope flows will also have this same type of essential singularity. This result then has very interesting implications to slope flows, and especially to interpreting field measurements.

Field observations taken along a slope could mistakenly be interpreted to support that a particular flow had become independent of slope length when in reality the flow was still undergoing its transient boundary layer development. In Sections 1.4 - 1.7 the conditions when a steady state slope flow becomes independent of slope length are thoroughly studied. The study of the transient slope flow introduces another caveat that investigators must understand before field measurements can be properly compared against theory.

Similar to the previous sections, the boundary layer is due to a temperature

difference between the ground and the air. In this case the temperature difference will be assumed to occur instantaneously. Such air/ground temperature differences do not really occur in nature, but the dynamics involved in creating a temperature difference by radiative heat transfer, or advection of an unequilibrated air mass over the slope, are generally fast compared to the development of the boundary layer. This assumption represents the most abrupt conditions which would start a boundary layer to develop. Field observations are likely to be less dramatic.

Method

The essence of the following method is the reduction of the coupled momentum and energy partial differential equations (PDE) to a single first order PDE describing the momentum boundary layer thickness as a function of time and position. The method of characteristics is applied to this equation, giving a simple technique to determine the extent of boundary layer development in a transient slope flow.

The transient problem is approached identically to the steady-state boundary layer solution given in Section 1.7. By analogy the following unsteady integral equations can be derived. The critical slope length parameter, λ , is assumed to be zero. Physically, this restricts the solutions to boundary layers which are steady state functions of the distance along the slope.

$$\frac{d}{dt} \int_0^{\delta(\chi,t)} U d\zeta + \frac{d}{d\chi} \int_0^{\delta} U(U - U_{\infty}) d\zeta + \frac{dU_{\infty}}{d\chi} \int_0^{\delta} U d\zeta = -Pr \frac{\partial U}{\partial \zeta} \Big|_{\zeta=0} + \int_0^{\delta} \Theta d\zeta \quad (1.8.1)$$

$$\frac{d}{dt} \int_0^{\delta_T(\chi,t)} \Theta d\zeta + \frac{d}{d\chi} \int_0^{\delta_T} U(\Theta - \Theta_{\infty}) d\zeta + \frac{d\Theta_{\infty}}{d\chi} \int_0^{\delta_T} U d\zeta = -\frac{\partial \Theta}{\partial \zeta} \Big|_{\zeta=0} \quad (1.8.2)$$

In the problem under consideration, the upper air is assumed to be quiescent and it remains at a constant temperature. Consequently, U_{∞} and Θ_{∞} are zero. This leads to Equations 1.8.3 and 1.8.4.

$$\frac{d}{dt} \int_0^{\delta(\chi,t)} U d\zeta + \frac{d}{d\chi} \int_0^{\delta} U^2 d\zeta = -Pr \frac{\partial U}{\partial \zeta} \Big|_{\zeta=0} + \int_0^{\delta} \Theta d\zeta \quad (1.8.3)$$

$$\frac{d}{dt} \int_0^{\delta_T(\chi,t)} \Theta d\zeta + \frac{d}{d\chi} \int_0^{\delta_T} U \Theta d\zeta = -\frac{\partial \Theta}{\partial \zeta} \Big|_{\zeta=0} \quad (1.8.4)$$

The technique of von Karman and Pohlhausen is employed next, thereby providing polynomial approximations for $U(\chi, \eta, t)$ and $\Theta(\chi, \eta, t)$ which satisfy the boundary conditions mentioned in Section 1.7.

$$U(\chi, \eta, t) = \frac{\delta^2(\chi, t)}{4Pr} \eta(1 - \eta)^2 \quad (1.8.5)$$

$$\Theta(\chi, \eta, t) = 1 - \frac{3}{2} \frac{\eta}{h} + \frac{1}{2} \left(\frac{\eta}{h} \right)^3 \quad (1.8.6)$$

In the above profiles, δ is the momentum boundary layer thickness, h is the ratio of the thermal boundary layer thickness to the momentum boundary layer thickness, and η is the dimensionless vertical height, $\frac{\xi}{\delta}$. The value of h is a constant. See Section 1.7 which describes this more completely.

The above profiles are then substituted into Equation 1.8.3 for momentum, and the integrals and partial derivatives are evaluated. This leaves the following partial differential equation for the boundary layer thickness.

$$\frac{\partial}{\partial t} \delta^2 + \frac{3}{14Pr} \frac{\partial}{\partial \chi} \delta^4 = 36Pr \left[\frac{3}{4} \left(1 - \frac{1}{h} \right) + \frac{1}{8h^3} \right] \quad (1.8.7)$$

This equation should be contrasted with the steady-state Equation 1.7.9. This equation is solved by the method of characteristics, bringing in the boundary condition $\delta(\chi = 0, t) = 0$ and the initial condition $\delta(\chi, t = 0) = 0$.

Results

The characteristic lines obtained are shown in Figure 1 on the following page. Two sets of lines are obtained; those beginning on the χ axis yield the value $\delta = [\frac{b}{a}\chi]^{\frac{1}{4}}$ and those beginning on the t axis $\delta = [bt]^{\frac{1}{2}}$, ($a = \frac{3}{14Pr}$, $b = 36Pr[\frac{3}{4}(1 - \frac{1}{h}) + \frac{1}{8h^3}]$, $h = 1.05$). The two regions overlap on the line $x = abt^2$.

Figure 2 shows the boundary layer height, $\delta(\chi)$, for various fixed times, generated by using Figure 1. This is the dimensionless plot times the lead coefficient of 2.21. The steady state solution obtained in Section 1.7 Equation 1.7.11 has a lead coefficient of 3.72. The difference in the two methods is a result of the extra polynomial which is introduced into the Equation 1.8.7 in the transient term. At the present there is no way to determine which of the two is most accurate. Lead coefficients excepted, functionally the two solutions are identical.

The essential singularity for the slope flow occurs at $U^*t = s$. The velocity, U^* , is the characteristic slope velocity defined in Equation 1.6.3a, t is time and s is the slope coordinate.

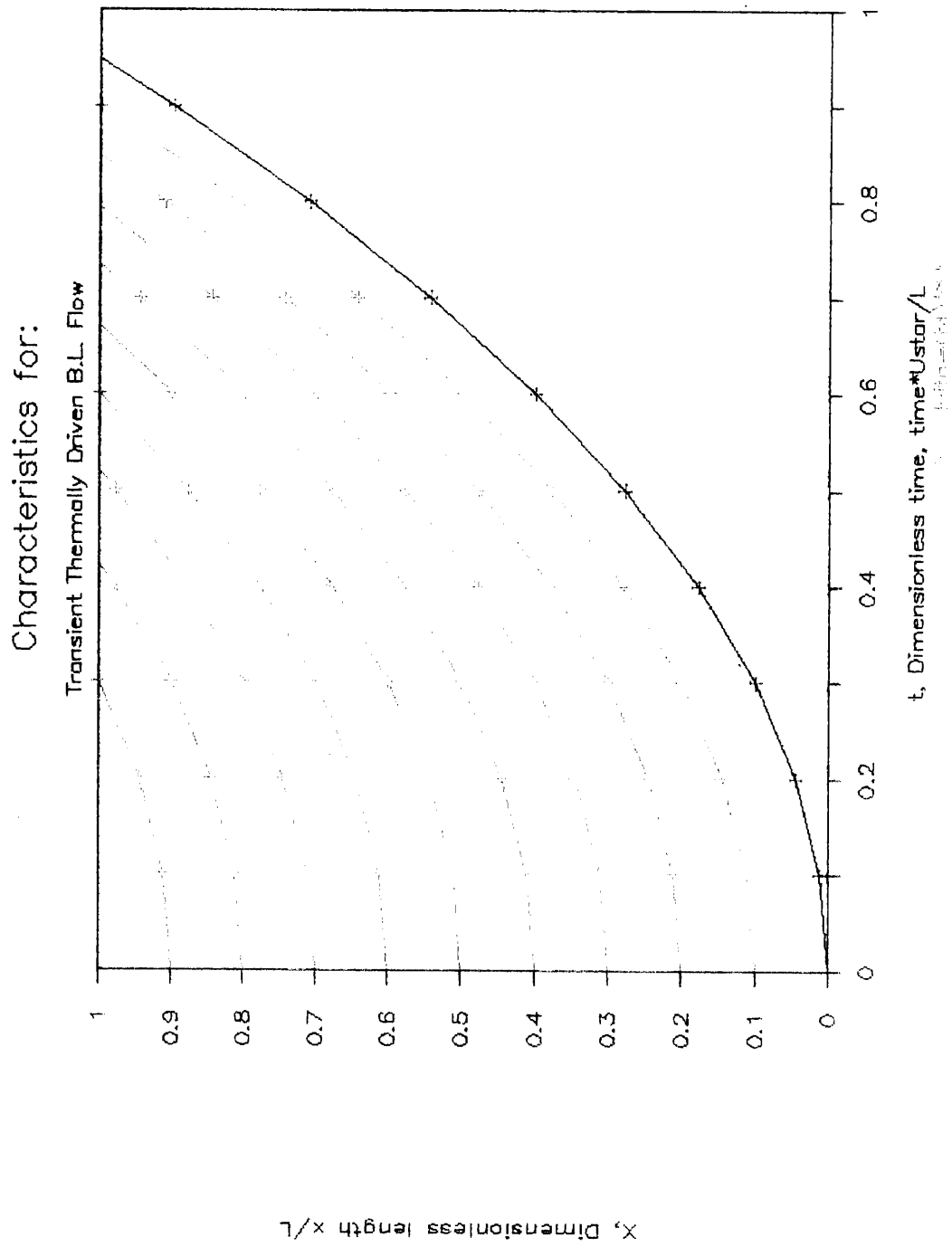


Figure 1. Characteristic lines for general transient slope flow.

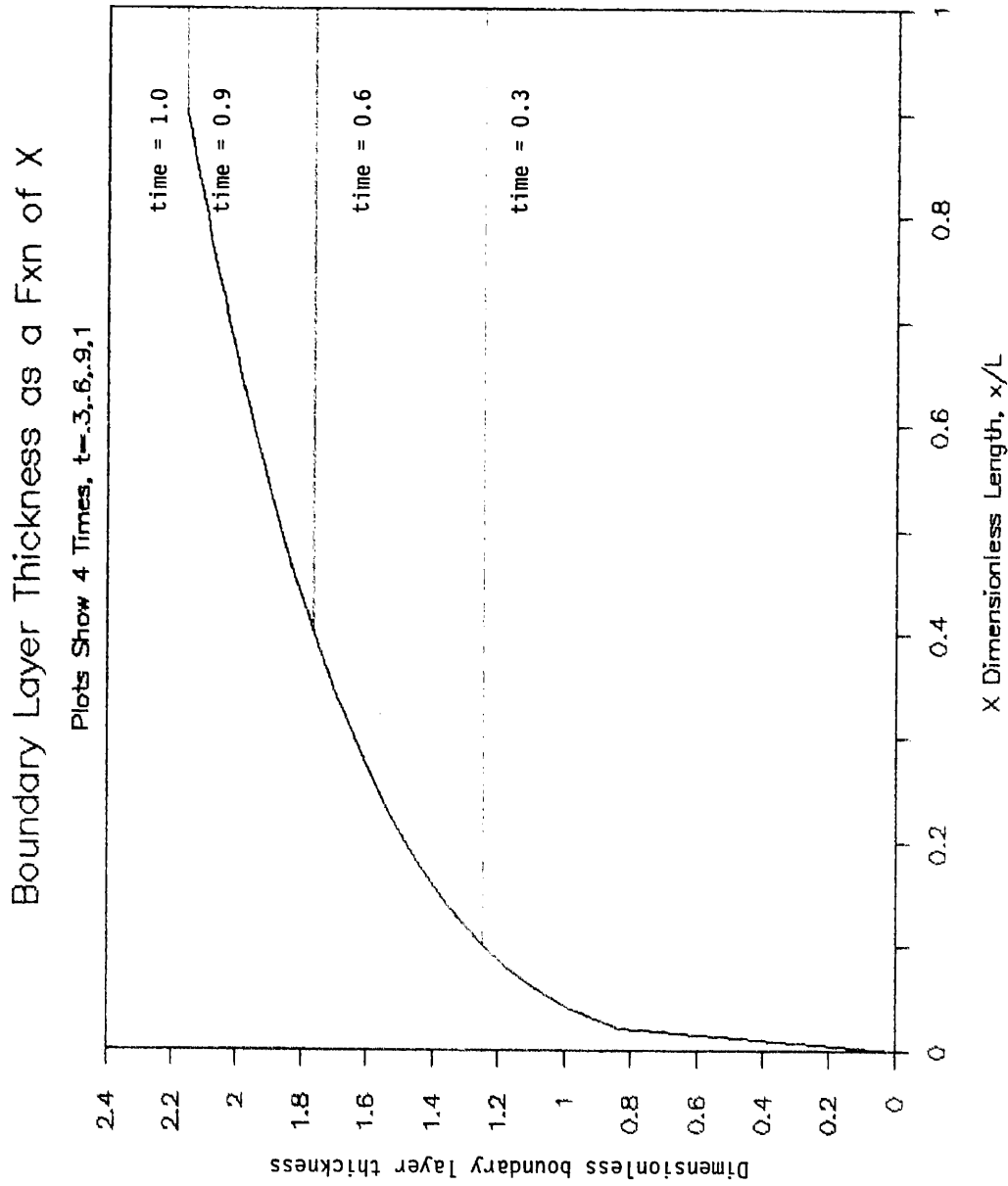


Figure 2. Boundary layer thickness as a function of dimensionless slope length and dimensionless time.

Conclusions

Four conclusions can be drawn from the above study of the transient flow.

1. The above method described herein provides a route for investigating systems of coupled PDE's derived from boundary layer analysis. Though approximate, the technique preserves a physical interpretation as to the origin of each term and does not require numerical techniques to complete the analysis. The next extension is to apply the technique to a more complete description of the problem discussed above.
2. Transient slope flows may be difficult to discern from steady slope flows which have boundary layers which have become independent of slope length.
3. Observations must be made with exceptional care so as to document every pertinent parameter to the flow as described in Section 1.6. Only then will it be possible to determine where along the slope of a given flow the boundary layer should become independent of slope length and where the observed flow was still experiencing transients.
4. When a high quality field study is done, these results should be compared with it and the existing models of other authors.

1.9 Katabatic Wind Along the Mountain Slopes

Bordering the Los Angeles Basin

Abstract

A theoretical framework is developed to interpret certain characteristics of the nighttime downsloping drainage winds from sloping terrain adjacent to a basin. The impact of the slope winds upon air pollution within a basin is discussed.

Specific data from ten South Coast Air Quality Management District (AQMD) Air Monitoring Sites along the base of the mountain slope which face the Los Angeles Basin were analyzed in order to determine the ground-to-ambient air differential temperature needed to generate the observed wind speeds.

Introduction

The mountains surrounding the Los Angeles Basin create conditions capable of producing light slope winds along their slopes. During the daytime, the influence of upslope winds within the basin is often obscured by the stronger sea breeze and larger scale dominant winds. However, at night and especially the early morning, these stronger winds subside and the local downslope winds may dominate. Downslope, or katabatic winds in the basin are low level parabolically shaped jets affecting the vertical region of air between the ground and about a hundred meters; typical temperature and velocity profiles are shown in Figure 1. Although these winds are generated on the mountain slopes, their impact is observed at the base of the mountains and many kilometers further into the basin as indicated by Figure 2. These nighttime drainage winds may be quite extensive and tend to reinforce each other along the perimeter of the basin. For example, the relatively straight east/west orientation of the San Gabriel and San Bernardino mountain ranges which bound the basin to the north causes downslope winds (from a northerly direction) from Pasadena to Riverside.

The significance of these pre-sunrise winds is their ability to alter the spatial distribution of the extensive morning automobile emissions and any other existing natural or anthropogenic gaseous pollutants throughout the basin. Katabatic winds alter the spatial pollutant distribution by three basic mechanisms: (1) new emissions are relocated from their source, (2) existing ambient emissions are pushed up and/or out of the way by the drainage, and (3) aged emissions or emissions from natural sources from higher elevations along the slope are transported down into populated areas. Furthermore, the stable stratification in the nighttime slope winds retains emissions near the ground because of reduced vertical mixing across the boundary layer.

Though difficult to separate from the sea breeze during the day, these winds

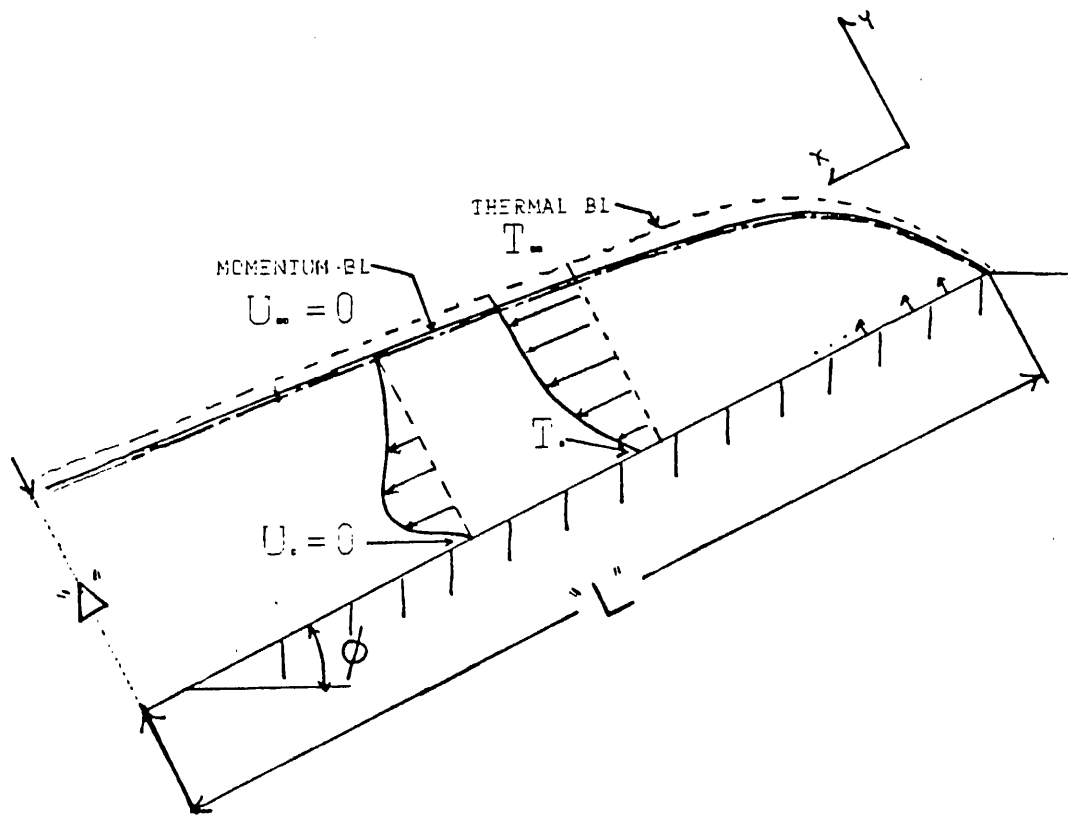


Figure 1. Velocity and temperature profiles in an idealized slope flow. The boundary layer thickness, Δ , the slope angle, ϕ , and slope length, L , are labeled.

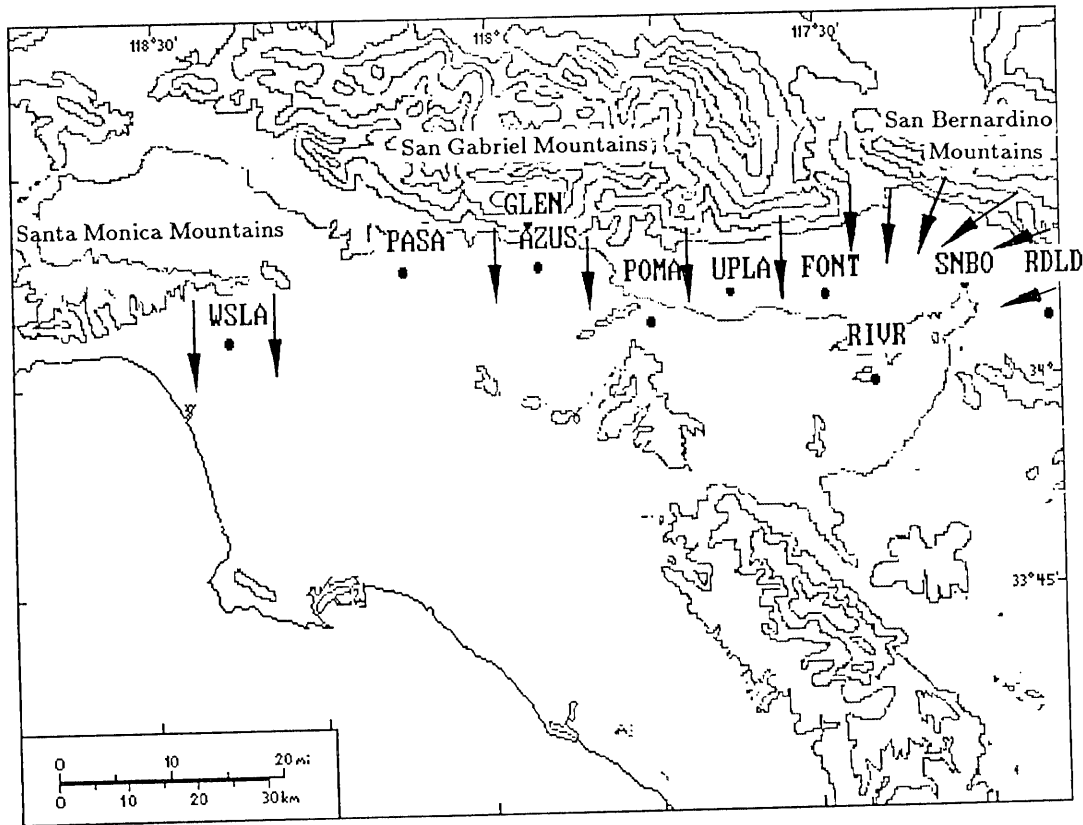


Figure 2. Slope wind measuring sites in the Los Angeles Basin. The arrows depict the nighttime downslope flow. The site abbreviations are defined in Table 1. The background lines are contour lines of elevation to highlight the location of the mountains in the basin.

may also affect the emission transport in local areas, causing variation in the wind field and mixing heights. Understanding the physics of the nighttime winds from a particular site may provide the means of analyzing the contributions of daytime slope winds at the same site.

In this paper two questions are addressed: (1) what is the nighttime ground to ambient air temperature difference which creates the buoyant conditions which drives the slope wind, and (2) why do observed nighttime slope winds from steep slopes tend to have lower speeds than their counter parts from gradual slopes?

This work was motivated primarily to provide a means to assess the particular wind fields generated by air flow simulations of the Los Angeles Basin, such as those from the Urban Airshed Model (see Chico, 1990). However, these two issues are relevant to explaining observations in any basin, both for up and downslope winds.

Method

The average wind speed, direction and associated standard deviations for each hour of the day at 31 wind monitoring sites in the basin were examined for the the summer of 1988. See Appendix A 2.3 of the section entitled “A Smog Season Examination of the LA-Glendale Divergence Zone.” From this set of 31 sites, the 10 sites bordering the San Gabriel, San Bernardino, and Santa Monica Mountains were selected. See Table 1, which also includes the average slope angle and slope length associated with the mountains near each site. The time period when each site was experiencing local winds, i.e., generated on the mesoscale, was determined from the plots in Appendix A 2.4 of the section entitled, “Examination of Wind Field Dynamics Using Cross Differencing.” From these data the hour of the day, the direction and the maximum speed of the wind during the period of local wind was calculated. See Table 2. The above mentioned data and the equations derived in the chapter entitled “A Comprehensive Dimensional Analysis of the Transport Properties of Atmospheric Slope Flows” are combined to estimate the maximum slope wind speed. From this estimate the differential temperature between the ground and the ambient air (above the slope wind layer) is calculated.

Site Label	Site Name	Adjacent Slope's Angle, ϕ (in degees)	Adjacent Slope's Length, L (in meters)
AZUS	Azusa	7.0	8000
FONT	Fontana	11.0	6000
GLEN	Glendora	7.7	4500
PASA	Pasadena	14.0	6000
POMA	Pomona	5.7	10000
RDLD	Redlands	3.8	30000
RIVR	Riverside	1.6	30000
SNBO	San Bernardino	3.8	30000
UPLA	Upland	13.2	10000
WSLA	West Los Angeles	2.5	7000

Table 1. Observation sites, and nearby slope's angle and length

Site Label	Hour of the Day (midnight = 0)	Max. Observed Avg. Speed, U_{Obs}^{Avg} (in m/s)	Observed Avg. Direction (degrees off north)
AZUS	5	0.9	351
FONT	4	0.5	15
GLEN	5	0.7	52
PASA	5	0.1	278
POMA	5	0.4	315
RDLD	4	0.5	108
RIVR	3	0.9	327
SNBO	4	0.4	98
UPLA	5	0.2	354
WSLA	3	0.5	348

Table 2. Nighttime slope/drainage wind characteristics

Estimation of the Maximum Slope Wind Speed

Wind data is measured at only one height above ground level, usually 9.1 meters (30 feet), at each of the sites. This presents a problem since without additional vertical wind data there is no direct way to determine how the measured speed compares to the maximum speed in the slope wind's vertical profile. However, an indirect method of determining the maximum speed from the measured speed at 9.1 meters has been devised. It proceeds through the following four steps: (1) the correct regime of the slope flow is determined from dimensional analysis, (2) the thickness of the slope flow layer is bounded, (3) the thickness is compared with the height of the measurement to see if it falls within the bounds, (4) given "(3)" is satisfied, the relationship between the maximum and measured speed is approximated. The method is developed in detail below.

To determine if the measurement height is within the slope flow boundary layer, and its relationship to the height of the actual speed maximum, the vertical thickness of the jet must be obtained. The most tenable method for obtaining this height, when it is not available by measurement, is to calculate it from its characteristic relationship to the slope parameters. To do this calculation correctly, the appropriate scaling regime for the slope flow must be determined.

The first step to determining the correct scaling regime for a slope flow is to compute the ratio of the slope's length, L , to its critical length, l_c . This ratio, defined to be

$$\lambda = \frac{\Gamma \sin \phi L}{\Theta_0 - \Theta_\infty} = \frac{L}{l_c}, \quad (1.9.1)$$

represents the most common demarkation point between regimes of atmospheric slope flows. In Equation 1 the parameters are as follows: the atmospheric lapse rate Γ , slope length L and angle ϕ , and the potential temperature difference $\Theta_0 - \Theta_\infty$. Practically, Equation 1.9.1 marks the distance, l_c , along the slope of

total length L , where the boundary layer ceases to increase in thickness. The characteristic and dimensionless parameters have different forms depending on the value of λ .

Determination of λ seems to require even more knowledge about the flow, i.e., the slope length and angle, the driving temperature difference, and the lapse rate. This turns out not to be true for several reasons, but before explaining why, consider a few facts about the parameters which make up λ . These facts may be helpful later in verifying some of the assumptions which are about to be made. In practice it can be difficult to determine L without extensive field measurement. The non-idealities in the slope surface and adequate knowledge of where the boundary layer begins along the slope being the main reasons. Frequently L is simply estimated (for downslope flows) by designating the apex of the slope as the leading edge (beginning of the boundary layer), then assigning L to the distance from there to a point along the slope. The uncertainties obtained by this approach are large, possibly 50 percent or more. (The actual boundary layer begins where a temperature difference exists, which is not necessarily the top of the slope). Next, the average potential temperature difference between the ground and the ambient air is the reason for this analysis, and if it were known this would be unnecessary. Third, the atmospheric lapse rate varies only in a small range so it presents no problems. The final parameter, the slope angle ϕ , is the key to finding λ , and mitigates the need to know the other three parameters accurately. The reason is the sinusoidal dependency of the critical length, l_c on ϕ , and the high accuracy to which ϕ can be known. Under atmospheric conditions, i.e., a wide range of lapse rates, slope lengths, and temperature differences, steep slopes (5 degrees or more) generally have $\lambda \geq 1$, and gradual slopes (1 degree or less) generally have $\lambda \geq 1$. The bounding mountains in the basin have steep slopes, and for this reason the entire analysis below is based on $\lambda \geq 1$.

The validity of this assumption is verified at a later point. Estimation of the Boundary Layer Thickness

When $\lambda \geq 1$ the relationship for the boundary layer thickness is as follows:

$$\Delta = Pr^{-\frac{1}{4}} \left[\frac{\alpha_T^2}{\Gamma(\sin^2\phi)g\beta} \right]^{\frac{1}{4}}. \quad (1.9.2)$$

To evaluate Equation 1.9.2 the following values were used. The eddy diffusivity, α_T , is estimated range from 1 to 3 m^2/s , typical for the neutral to stable nighttime conditions in the basin. The lapse rate is 0.01 (*degrees K/m*) which is an upper bound for this parameter. The slope angles for the cases being considered range from 1.6 to 14 degrees. The average summertime surface temperature (T) during the early morning hours is between 285 - 290 K. The thermal coefficient of volumetric expansion, $\beta = 1/T$, comes from the ideal gas law. The gravitational constant is 9.8 m/s . The Prandtl number is assumed to be of order 1. For these conditions the calculated boundary layer thickness falls between 18 and 96 meters. The wind profile due to a buoyantly driven layer is roughly parabolic, with a maximum speed approximately one third of the boundary layer up from the ground (Shair, 1987). Taking a third of this gives an estimated maximum height of from 6 to 32 meters, which bounds the height at which measurements are being taken.

Even though the wind measurements are being taken inside the slope wind boundary layer, the probability that the maximum speed is observed in any measurement is low. Thus, a reasonable assumption is the average of the observed speeds, at a site, is the average speed of the vertical boundary layer. This takes into account changes in the height of the maximum speed from measurement to measurement. The maximum boundary layer speed, U_{Obs}^{Max} , has been determined to be 1.77 times the average speed, U_{Obs}^{Avg} , by means of the approximate Von Karman-Pohlhausen solution for buoyant flows. The maximum observed speed,

U_{Obs}^{Max} , has already been shown to be 0.625 times the characteristic velocity, U^* , under a wide range of conditions. Based on the above analysis, the wind speed that is measured at each site will be assumed to be the vertical average slope wind speed. Consequently, the characteristic velocity for the flow will be calculated as

$$U^* = 2.84U_{Obs}^{Avg}. \quad (1.9.3)$$

Without additional data it is inappropriate to refine the estimate of U^* further.

Determination of the Differential Temperature

The temperature driving force is determined by rearranging the characteristic equation for the slope wind velocity given below. The characteristic velocity for slope wind for $\lambda \geq 1$ is related to the slope and temperature parameters by Equation 1.9.4.

$$U^* = Pr^{-\frac{1}{2}} \left[\frac{g\beta|\theta_0 - \theta_\infty|^2}{\Gamma} \right]^{\frac{1}{2}} \quad (1.9.4)$$

Rearranging Equation 1.9.4 and combining it with Equation 1.9.3 provides a means of estimating the average differential temperature for the slope flows at each site.

$$|\theta_0 - \theta_\infty| = 2.84Pr^{\frac{1}{2}} \left[\frac{\Gamma}{g\beta} \right]^{\frac{1}{2}} U_{Obs}^{Avg} \quad (1.9.5)$$

For the calculations described in the Results section, the parameters Pr , g , β , and Γ have the values given above. The values for U^* were obtained from Table 2.

Slope Wind Speed and Slope Length

A graph, Figure 3, is created to show the relationship between the characteristic slope wind velocity and slope length (for $\lambda \leq 1$) which also conveys the relationship between slope angle and the critical slope length. Thus the transition to $\lambda \geq 1$ is also made clear for a wide range of slope angles. The characteristic slope wind velocity for $\lambda \leq 1$,

$$U^* = Pr^{-\frac{1}{2}} [g(\sin \phi)\beta|\theta_0 - \theta_\infty|L]^{\frac{1}{2}} = [Pr^{-1}\sigma^2gL]^{\frac{1}{2}} \quad (1.9.6)$$

is plotted as lines of constant slope strength, $\sigma = [(\sin \phi)\beta|\theta_0 - \theta_\infty|]^{\frac{1}{2}}$. Three lines of constant slope angle, ϕ , 25, 5, and 1 degree(s), are plotted on top of the first plot, such that their intersection points with the lines of constant slope strength are the critical slope length, l_c . This graph is general and can be applied under the full range of atmospheric conditions worldwide. (It has even been applied when diagnosing the potential for slope flow in the polar regions of Martian atmosphere).

From Figure 3 it is clear that the same steep slopes responsible for a large slope strength also create conditions which hasten reaching a maximum speed along the slope. This is consistent with observations of katabatic winds worldwide, i.e., the shallow slopes of Antarctica generate strong katabatic winds and the steep slopes of mountains exhibit gentle breezes.

Slope Wind Speed vs. Slope Length

Atmospherically Observable Conditions

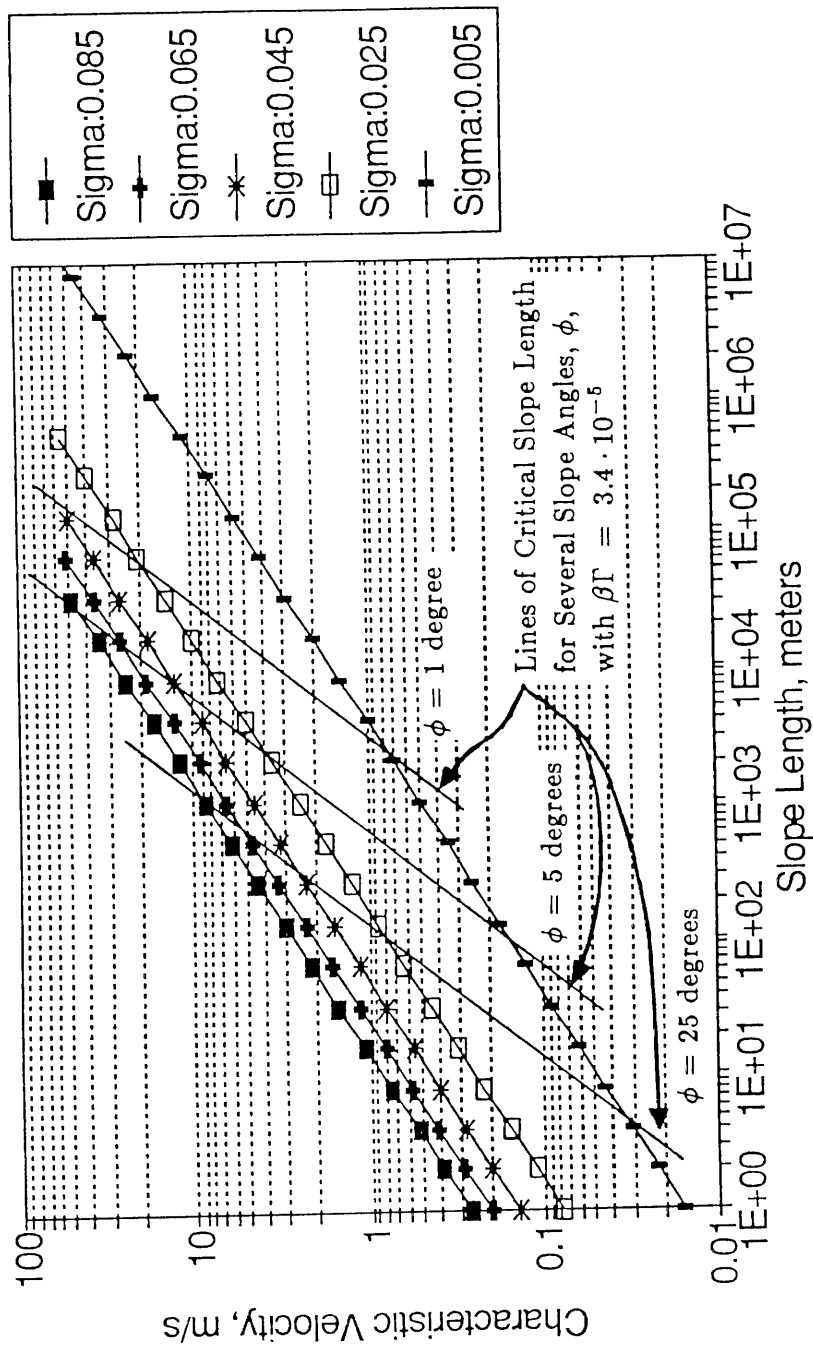


Figure 3. Characteristic slope wind velocity as a function of slope length and slope strength also showing lines of critical slope length for different slope angles.

Results

Temperature Differential

The ground-to-ambient-air temperature difference for the 10 sites, as computed from Equation 1.9.5, are compiled in Table 3. The values range from a 0.1 to 1.5 K, which is consistent with the isolated measurements made at Scholl Canyon by Shair; refer to Table 3 in the chapter entitled "A Comprehensive Dimensional Analysis of the Transport Properties of Atmospheric Slope Flows." With the temperature difference known, the critical length and λ can be determined from equation 1. The values obtained verify the original assumption that λ was in excess of 1.

Spatial Distribution of Driving Forces

Nine of the ten sites experience drainage from katabatic winds on the average. Pasadena is the only anomalous site, i.e., its average wind direction during the local wind period does not flow perpendicular to the nearby mountain range. But when one notes the slopes near Pasadena are among the steepest of those studied, and the potential for slope wind is extremely small, it is not surprising slope wind is not observed.

Quick Estimation of Slope Wind Speed

Figure 3 can be used to rapidly estimate slope flow speeds worldwide. To use the graph, compute a slope strength, σ . Using this slope strength and the slope angle, find their intersection on the graph. The velocity at this point is the maximum for that slope under the chosen conditions, the slope length at this point is l_c . The wind speed does not increase with increasing slope length beyond this point. If the actual slope length is less than l_c , then follow the line of constant slope strength down and to the left to this slope length, and read off the characteristic velocity. The parameters $\beta\Gamma = 3.4 \cdot 10^{-5}$ were picked to

Site Label	Temp. Diff. $ \theta_0 - \theta_\infty $ (degrees K)	Critical Slope Length, l_c (in meters)	λ
AZUS	1.37	1100	7.3
FONT	0.77	400	15
GLEN	1.07	800	5.6
PASA	0.15	70	85
POMA	0.61	600	17
RDLD	0.77	1100	27
RIVR	1.37	4900	6.1
SNBO	0.61	900	33
UPLA	0.31	90	110
WSLA	0.77	1100	6.3

Table 3. Nighttime differential temperature, critical slope length, and dimensionless slope parameter

be average atmospheric conditions. If the values of these parameters are known very accurately for particular circumstances, it might be desirable to recompute the lines of constant ϕ . Under most conditions these lines will not be in error by more than a few percent.

If the slopes of the Los Angeles Basin were not subject to the limiting physical restrictions associated with $\lambda \geq 1$, they would be capable of generating downslope wind speeds in excess of 10 *m/s*. However, because these restrictions do exist, observations in the basin are restricted to the lower left regions of the plot in Figure 3. In addition, the observations presented herein support the relationship between steeper slopes having slower observed katabatic winds.

Conclusions

The critical slope length, l_c , may be more aptly called the effective slope length for steep slopes, since it really is the maximum segment of the total slope length which must be experiencing a potential temperature difference in order to create a flow. For steep slopes, the place along the slope where the katabatic wind develops may be of little importance because the critical slope length is short compared to the total slope length. For the slopes which surround the Los Angeles basin, the critical length is modest compared to the total slope lengths, 20 percent or less in all cases studied. In terms of the computation interval of numerical models, this means energy and momentum calculations along slopes should be done at intervals of 0.5 km at the most. The small temperature differences which drive the flow, on the order of a degree, and the vertical bound on the slope flow thickness of 100 meters means vertical computational resolution on the order of 30 meters or less. It is clear that considerable computational power is required to reproduce these flows accurately above any large surfaces.

Though no documented evidence is known by the author to support this claim, it is suggested that the intrusion of a large amount of stable slope air centering around Glendora and Azusa may on certain mornings initially slow the transport of western basin emissions eastward. Furthermore, the emissions from the late afternoon and evening of the previous day may be revisiting the basin by means of the slope flow drainage through Azusa and Glendora. The detailed physics of the flow from this area could be studied quite effectively using atmospheric tracers. The knowledge obtained would benefit both the conceptual model and the numerical calculations at this and other locations.

**A 1.9 Appendix : Average Wind Speed and Direction for
the Summer of 1988 at 31 AQMD Wind Monitoring Sites**

Introduction

A number of the experiments and studies described in this thesis benefit from having a seasonal average wind speed and direction for each hour of the day at each of the 31 Air Quality Management District wind monitoring sites. The data from this composite average day can be used directly, as it is in this paper where the data from individual days would not suffice, or it can be used as a reference, as it is in later sections where the average summertime day is used to make comparisons to individual day's measurements.

The year 1988 was chosen for many reasons. One, the data was current. Two, the quality of the data could be checked through direct contact with the AQMD personnel currently responsible for it. Three, 1988 was a good year for ozone observations relative to other recent years, which is important for later uses. Only the data from one summer was used to compute the average because it eliminated possible year to year variations.

Averaging Method

Wind data was obtained from the South Coast Air Quality Management District's Telemetry System from May to September of 1988. This data was measured at 31 sites, on an hourly basis, and consists of vector-averaged wind speed and direction. The data is put into binary files consisting of one month blocks, i.e., the WIND format as described by the WIND Manual Version 2.0.

The data was averaged using the special purpose program AVEWIND, designed to accept WIND formatted data. The data averaging algorithm used by AVEWIND converts the wind speed and direction data into its vector components, performs the vector averaging and then converts the data back to its original speed and direction format. Standard deviations of the speed and direction at each site and hour are also computed by AVEWIND. These are computed from the average speed and direction. In the calculation of the wind direction standard deviation, angle differences are limited to 180 degrees.

All hours are given in Pacific Daylight Savings Time (PDT). Hour 0 is midnight (PDT), and corresponds to the data gathered from midnight up to hour 1, and so on for the remaining 23 hours.

Averaged Data

Tables 1 - 4 show the results of the averaging and the standard deviations for wind speed and direction. These results will be used and referred to throughout this thesis.

The four data sets are useful for a variety of reasons, including characterizing each site on average, and providing a means of interpreting the averages based on their uncertainty. Similarities and differences in the sites based on geographical location can also be examined using Tables 1 - 4.

Figures 1 - 10 were compiled to provide a qualitative and quantitative method by which the sites from similar geographic areas could be compared on the basis of their wind direction standard deviations. These plots provide a measure of both the site and the regional uncertainties in the wind direction at every hour.

The wind direction standard deviation plots provide a unique method of assessing the influence of certain wind patterns in a given region. To see how, consider the following example. The San Fernando Convergence zone may move across three wind monitoring sites in the Los Angeles Basin: Burbank, Reseda, and Newhall in the afternoon of any day. The zone is known to move from west to east, often impacting Reseda, frequently passing Newhall, and seldom reaching Burbank. When the divergence zone passes a site, the wind direction changes by roughly 120 degrees. Figure 1 clearly demonstrates how the east to west migration of the zone increases the uncertainty in: (1) Reseda's wind direction at about 15 PDT, (2) Newhall's wind direction two hours later, and (3) Burbank's wind direction slightly another two hours later.

Similar inferences may be made from the other figures, and using the other three sets of data.

HOUR	0	1	2	3	4	5	6	7	8	9	10	11
STATION												
CELA	1.1	0.5	0.4	0.7	0.6	0.9	0.5	0.5	1.1	1.9	2.8	3.8
WSLA	0.2	0.2	0.1	0.5	0.3	0.3	0.0	0.9	1.8	3.2	5.1	6.5
HAWT	1.4	1.2	0.8	0.5	0.6	0.6	0.4	0.7	1.1	2.6	4.8	6.9
LGBH	0.3	0.6	0.4	0.4	0.6	0.5	0.8	0.9	1.2	1.9	2.5	3.5
WHIT	0.3	0.1	0.1	0.1	0.2	0.3	0.3	0.3	0.5	1.0	1.2	2.0
RESE	0.3	0.2	0.1	0.2	0.2	0.3	0.2	0.2	0.4	0.8	0.8	1.0
BURK	1.9	1.6	1.2	1.0	0.6	0.9	0.8	1.0	1.1	1.9	2.6	3.1
PASA	0.1	0.1	0.1	0.0	0.1	0.1	0.0	0.2	0.3	0.6	0.8	1.1
AZUS	1.6	1.4	0.8	0.8	0.5	0.9	0.7	0.3	1.2	1.9	3.1	3.9
GLEN	0.3	0.3	0.4	0.5	0.6	0.7	0.6	0.2	0.6	1.4	2.1	2.7
POMA	0.9	0.8	0.6	0.5	0.4	0.4	0.4	0.3	0.7	1.1	1.8	2.8
PICO	1.9	1.1	1.0	0.9	0.7	0.5	0.8	1.1	1.5	2.8	3.9	4.7
LYNN	1.3	0.9	0.7	0.6	0.6	0.6	0.7	0.7	1.0	1.5	2.3	4.0
NEWL	0.1	0.3	0.3	0.2	0.0	0.1	0.0	0.1	0.1	0.7	2.3	4.2
LANC	1.5	1.2	1.2	0.9	1.0	0.8	0.7	0.7	1.1	1.1	1.1	0.7
LAHB	0.5	0.5	0.4	0.4	0.2	0.1	0.2	0.5	0.8	1.5	2.6	3.2
ANAH	1.1	1.0	0.9	0.8	1.0	0.7	0.9	0.9	1.4	1.9	2.8	3.8
LSAL	0.6	0.5	0.5	0.4	0.3	0.4	0.4	0.6	0.8	1.5	2.7	4.1
COST	0.4	0.3	0.2	0.1	0.2	0.3	0.3	0.4	0.8	1.6	2.2	3.0
TORO	0.9	0.6	0.6	0.7	0.6	0.6	0.8	0.8	0.8	0.8	1.1	1.4
NORC	0.7	0.7	0.5	0.4	0.4	0.3	0.5	0.5	1.1	1.3	2.6	3.8
RIVR	1.4	1.0	0.8	0.9	0.4	0.3	0.5	0.7	0.6	1.3	2.6	4.4
PERI	0.0	0.2	0.1	0.0	0.1	0.1	0.2	0.5	0.6	0.5	0.5	1.2
HEME	1.4	1.1	0.7	0.9	0.7	0.6	0.5	0.4	1.0	1.5	2.1	2.9
BANN	5.1	5.1	4.7	4.3	4.3	4.1	4.1	3.9	3.8	4.0	4.2	4.9
PLSP	6.4	5.8	5.0	4.7	4.2	3.7	3.3	2.8	1.5	0.7	0.2	0.0
UPLA	0.5	0.4	0.3	0.3	0.2	0.2	0.2	0.2	0.6	1.0	1.5	2.0
FONT	0.4	0.4	0.2	0.3	0.5	0.3	0.4	0.3	0.5	1.7	2.3	3.3
SNBO	1.3	0.9	0.8	0.5	0.4	0.2	0.4	0.4	0.5	1.0	2.0	2.7
RDLD	0.1	0.3	0.2	0.4	0.5	0.4	0.2	0.1	0.5	0.8	1.9	2.2
CRES	1.5	1.4	1.4	1.0	1.0	0.9	1.1	1.4	2.4	3.2	3.3	3.2
HOUR	12	13	14	15	16	17	18	19	20	21	22	23
STATION												
CELA	5.8	7.4	9.0	8.9	9.0	8.3	6.8	5.7	3.8	2.7	2.0	1.5
WSLA	7.5	7.5	7.9	7.6	7.2	6.3	4.7	3.0	1.7	0.9	0.7	0.3
HAWT	8.6	9.9	10.0	9.9	9.1	8.7	7.5	6.1	5.1	3.7	3.0	2.3
LGBH	4.1	4.3	4.6	4.8	5.8	5.1	4.7	3.9	3.0	1.8	0.9	0.5
WHIT	2.9	4.3	4.5	5.2	4.2	4.0	3.3	2.5	1.8	1.0	0.6	0.2
RESE	1.7	1.9	2.1	1.5	0.2	0.1	0.7	0.5	0.5	0.4	0.3	0.4
BURK	3.1	3.6	3.8	3.6	3.7	3.4	3.2	3.4	3.0	2.6	2.6	2.4
PASA	1.6	2.3	2.4	2.9	2.2	2.1	1.5	0.7	0.3	0.2	0.0	0.1
AZUS	4.6	5.9	6.5	7.0	7.3	6.6	6.3	5.3	4.0	2.9	2.3	1.8
GLEN	2.9	3.6	4.2	4.8	4.2	3.7	2.5	1.2	0.6	0.3	0.1	0.1
POMA	3.6	4.0	5.2	5.4	5.1	5.1	4.2	3.5	2.6	1.9	1.5	1.2
PICO	5.4	7.1	8.0	8.4	8.2	7.5	6.9	5.5	4.7	3.4	2.8	2.4
LYNN	6.1	7.4	8.2	8.4	8.4	7.8	7.0	6.0	4.4	3.4	2.5	1.6
NEWL	4.9	4.6	4.7	4.8	4.2	3.0	1.8	1.3	0.5	0.4	0.3	0.3
LANC	1.3	2.2	3.5	3.9	4.4	3.9	3.8	3.6	2.8	1.9	1.7	1.9
LAHB	4.4	5.4	6.0	5.9	5.9	5.2	4.8	4.1	3.0	2.1	1.5	0.8
ANAH	5.4	6.1	6.6	6.7	6.3	6.0	4.9	4.2	3.6	2.4	1.8	1.6
LSAL	4.9	5.3	5.8	5.1	4.5	3.9	3.1	2.1	1.7	1.1	0.7	0.6
COST	3.6	4.0	4.0	3.6	3.5	2.9	2.4	1.7	1.2	0.8	0.7	0.5
TORO	2.3	2.9	3.0	2.9	2.3	2.6	1.8	1.2	1.0	0.6	0.7	0.8
NORC	6.7	9.1	10.6	11.4	10.9	9.9	8.5	6.1	4.0	2.6	1.7	1.1
RIVR	5.9	7.6	8.9	10.4	10.8	9.6	7.2	4.6	3.5	2.6	2.3	1.6
PERI	2.7	3.8	5.5	6.1	7.3	6.3	4.0	1.2	0.6	0.3	0.2	0.3
HEME	4.2	5.1	5.4	7.0	7.3	7.2	6.2	4.7	3.6	2.6	2.6	1.8
BANN	4.7	4.8	5.2	6.5	7.1	8.0	7.9	7.8	6.8	5.9	5.2	5.2
PLSP	0.6	1.0	1.5	2.3	3.5	4.8	6.3	8.4	9.4	8.8	8.4	7.6
UPLA	2.5	2.8	3.1	3.4	3.4	3.2	2.9	2.6	1.9	1.5	0.9	0.6
FONT	4.3	5.3	6.4	8.0	8.2	7.7	6.6	4.4	3.2	1.6	0.9	0.6
SNBO	3.7	4.5	4.8	6.9	8.1	7.9	7.2	5.4	4.2	3.2	2.3	1.7
RDLD	3.1	3.6	4.3	4.9	5.0	4.6	3.4	2.0	1.2	0.6	0.2	0.2
CRES	3.8	3.8	4.1	4.0	4.3	3.8	3.1	2.5	2.2	1.9	1.5	1.5

Table 1. Average wind speed from 31 SCAQMD Telemetry System Sites for the period May - September 1988 at all 24 hours. Data is in units of miles per hour.

HOUR STATION	0	1	2	3	4	5	6	7	8	9	10	11
CELA	2.6	3.2	3.1	3.0	2.6	2.8	3.2	3.2	2.8	2.9	2.6	3.3
WSLA	2.4	2.8	2.3	2.2	2.4	2.3	2.8	2.5	2.4	3.0	3.4	3.6
HAWT	2.8	2.9	2.5	2.9	2.7	2.3	2.6	2.8	3.0	3.2	3.1	3.4
LGBH	2.9	2.4	2.3	2.5	2.5	2.5	2.2	2.5	2.7	2.7	2.7	2.6
WHIT	2.0	1.9	1.9	1.8	1.7	1.5	2.0	2.1	2.2	2.7	2.5	2.5
RESE	1.9	1.9	2.1	1.6	1.8	1.8	1.8	2.1	2.2	2.4	2.5	3.5
BURK	2.1	2.4	1.9	2.0	1.5	1.8	1.7	2.1	2.4	2.5	2.5	2.6
PASA	0.7	0.6	0.5	0.5	0.5	0.6	0.8	0.9	1.0	1.4	1.5	1.7
AZUS	2.1	2.0	2.2	2.3	2.5	2.3	2.4	2.6	2.1	2.4	2.2	2.4
GLEN	0.9	1.0	1.0	1.0	1.2	1.2	1.0	1.1	1.1	1.5	1.8	2.0
POMA	1.2	1.4	1.1	1.0	1.1	0.9	1.1	1.1	1.5	1.7	2.3	2.4
PICO	2.4	2.3	2.0	2.1	1.9	2.4	2.4	2.8	2.6	3.2	2.9	3.1
LYNN	2.5	2.3	2.1	2.1	2.0	1.9	2.2	2.3	2.5	2.6	3.1	3.3
NEWL	2.2	1.7	1.3	1.6	1.5	1.0	1.3	2.0	2.5	3.5	3.8	4.0
LANC	2.2	2.5	2.4	2.0	2.5	2.0	1.9	2.5	3.0	3.2	4.3	4.7
LAHB	2.3	2.2	2.1	2.2	2.2	2.3	2.4	2.5	2.4	3.0	2.7	2.7
ANAH	2.0	2.0	2.0	2.1	2.2	2.2	2.5	2.6	2.4	2.6	2.7	2.8
LSAL	2.5	2.5	2.2	2.2	2.2	2.1	2.3	2.5	2.8	2.8	3.3	3.6
COST	1.7	1.5	1.4	1.5	1.2	1.2	1.9	1.5	1.8	2.1	2.5	2.7
TORO	1.8	1.4	1.9	1.5	1.6	1.7	1.8	2.0	2.7	2.6	3.2	3.3
NORC	2.4	2.3	2.1	1.8	2.0	1.9	2.2	2.2	2.5	3.7	3.8	4.6
RIVR	3.0	2.8	2.9	3.0	2.9	2.8	3.0	5.7	3.0	4.0	3.8	3.6
PERI	1.5	1.6	1.3	1.4	1.2	1.3	1.6	2.2	2.2	3.0	3.5	4.3
HEME	2.2	2.1	1.8	2.2	1.9	1.7	2.1	2.1	2.5	2.6	2.9	3.5
BANN	3.1	3.5	3.3	3.2	3.4	3.4	3.8	4.2	5.5	6.1	6.7	6.4
PLSP	4.3	4.6	4.5	4.8	4.9	4.2	4.1	5.1	5.2	5.1	5.9	6.7
UPLA	1.0	0.8	0.8	0.9	0.8	0.9	0.9	0.9	1.0	1.2	1.2	1.2
FONT	1.7	2.2	2.0	2.0	2.0	1.9	2.6	3.2	2.5	2.8	3.5	3.3
SNBO	2.4	1.9	1.6	1.4	1.5	1.8	1.7	1.8	2.3	3.2	3.3	3.2
RDLD	1.4	1.3	1.3	1.2	1.0	1.0	1.3	1.6	2.0	2.3	2.4	2.4
CRES	3.2	3.0	3.1	3.0	2.5	3.0	3.3	3.3	3.4	3.6	4.0	4.1
HOUR STATION	12	13	14	15	16	17	18	19	20	21	22	23
CELA	3.3	3.8	3.6	3.6	3.2	3.2	3.1	3.1	2.8	2.6	2.3	2.5
WSLA	3.4	3.0	3.2	3.0	3.0	2.5	2.4	2.7	2.5	2.6	2.5	2.4
HAWT	3.1	3.6	3.0	3.0	3.1	3.3	3.4	3.1	3.4	3.0	3.0	3.0
LGBH	3.1	3.6	3.9	3.4	3.5	3.5	2.9	2.7	2.4	2.6	2.3	2.4
WHIT	3.0	3.2	3.1	3.4	2.9	3.2	2.9	2.4	2.6	2.5	2.0	2.0
RESE	2.7	3.3	3.2	4.0	5.2	5.0	4.2	3.2	2.6	2.3	1.7	1.5
BURK	3.0	3.0	2.9	3.0	2.9	3.2	2.6	2.6	2.5	2.4	2.0	2.1
PASA	1.9	2.1	2.1	2.3	2.2	1.7	1.6	1.3	0.8	0.5	0.9	0.7
AZUS	2.7	2.5	2.8	2.6	2.2	2.1	2.0	2.2	2.1	2.1	2.0	1.9
GLEN	2.4	2.6	2.7	2.6	2.4	2.5	2.1	1.6	1.6	1.4	1.1	1.1
POMA	2.7	2.9	3.1	2.7	2.7	2.7	2.6	2.3	2.2	1.9	1.8	1.5
PICO	3.4	3.7	3.3	2.8	3.4	2.7	2.6	2.4	2.7	2.3	2.3	2.5
LYNN	3.2	3.4	3.3	3.2	3.0	3.1	2.3	2.6	2.3	2.2	2.3	2.5
NEWL	4.0	3.8	3.3	3.6	3.8	3.5	3.3	2.7	2.1	1.8	1.5	1.5
LANC	4.9	4.8	4.3	4.1	4.2	4.2	3.5	3.2	2.7	2.5	2.1	2.4
LAHB	2.9	2.9	3.0	2.8	3.0	2.8	2.6	2.6	2.4	2.7	2.4	2.3
ANAH	2.9	2.5	2.8	2.6	2.8	2.4	2.4	2.3	2.2	2.5	2.2	2.1
LSAL	4.0	4.2	4.8	4.5	4.1	3.8	3.5	2.9	2.5	2.2	1.9	1.7
COST	3.0	3.3	3.0	3.0	2.9	2.8	2.4	2.1	1.8	1.6	1.6	1.3
TORO	3.7	3.7	3.9	3.6	3.5	3.1	2.8	2.4	2.0	1.8	1.6	1.7
NORC	4.9	4.6	4.6	4.3	4.2	3.9	3.9	3.2	3.0	2.9	2.8	2.5
RIVR	3.8	4.3	4.3	4.9	4.8	4.2	3.7	3.1	2.9	2.6	2.5	2.4
PERI	4.5	4.7	4.5	4.5	4.7	4.3	4.3	4.5	3.8	3.2	3.0	2.6
HEME	3.7	4.2	4.1	3.7	3.5	2.7	2.6	2.0	2.0	2.1	2.3	2.3
BANN	6.5	6.3	5.7	5.2	4.8	5.0	4.5	4.2	3.4	3.6	3.3	3.4
PLSP	7.1	7.4	7.3	6.6	6.2	5.9	5.6	5.8	6.1	5.5	5.4	5.2
UPLA	1.2	1.4	1.2	1.3	1.3	1.2	1.2	1.1	1.1	1.2	1.1	1.1
FONT	3.0	3.6	3.4	6.3	3.4	3.1	2.9	2.5	5.3	2.2	2.0	1.9
SNBO	3.4	3.6	3.9	3.9	3.7	3.5	2.8	2.8	2.4	2.1	1.9	1.9
RDLD	2.6	3.0	3.0	3.1	2.8	2.5	2.4	1.9	1.7	1.3	1.6	1.2
CRES	3.8	4.5	3.9	4.0	3.5	3.7	3.7	3.6	3.4	3.4	3.5	3.8

Table 2. Standard deviation of the wind speed from 31 SCAQMD
Telemetry System Sites for the period May - September
1988 at all 24 hours. Data is in units of miles per hour.

HOUR	0	1	2	3	4	5	6	7	8	9	10	11
STATION												
CELA	253	259	311	346	3	30	28	143	185	200	197	213
WSLA	333	341	356	348	332	344	327	181	190	211	219	222
HAWT	235	218	204	186	147	155	118	163	184	230	251	257
LGBH	298	238	178	70	80	95	107	114	163	184	191	200
WHIT	258	59	46	27	44	65	61	97	183	211	202	216
RESE	105	107	125	129	143	152	189	162	122	109	112	125
BURK	144	151	170	177	196	196	202	194	180	164	145	151
PASA	228	240	258	273	270	278	247	163	159	184	185	196
AZUS	264	275	280	298	318	351	352	258	232	228	229	233
GLEN	42	37	40	42	48	52	61	175	203	222	225	232
POMA	306	301	310	321	315	321	302	234	231	251	270	279
PICO	195	189	182	167	161	167	209	229	219	211	221	220
LYNN	249	244	226	223	216	198	194	196	199	204	210	226
NEWL	103	107	121	111	107	111	85	1	306	293	176	164
LANC	188	187	172	169	167	169	184	190	198	190	198	178
LAHB	242	280	329	354	73	29	138	200	250	249	246	249
ANAH	195	186	178	162	162	140	131	162	187	202	217	220
LSAL	278	315	308	342	17	358	13	19	302	264	255	251
COST	244	245	285	2	18	33	32	12	206	190	208	209
TORO	121	118	121	114	116	106	115	153	177	218	248	256
NORC	224	224	216	211	191	192	196	226	318	326	294	290
RIVR	318	318	326	327	314	313	281	202	215	254	271	290
PERI	317	113	77	117	133	194	151	123	130	83	49	16
HEME	186	183	179	176	160	166	191	288	274	265	255	248
BANN	265	264	264	265	275	273	263	253	247	235	226	223
PLSP	317	319	316	315	314	314	312	297	224	188	151	148
UPLA	315	326	336	343	350	354	354	149	184	217	226	240
FONT	315	348	354	0	15	12	18	157	194	219	235	233
SNBO	194	193	151	141	98	53	83	101	183	215	218	216
RDLA	99	103	108	108	108	110	108	343	297	289	282	285
CRES	251	254	243	257	257	255	246	204	148	152	167	187
HOUR	12	13	14	15	16	17	18	19	20	21	22	23
STATION												
CELA	226	238	244	247	249	253	249	249	242	234	229	237
WSLA	233	238	238	243	240	240	234	228	213	209	210	292
HAWT	260	262	264	271	270	269	272	267	262	254	245	246
LGBH	210	222	236	248	260	270	269	273	273	281	278	285
WHIT	216	212	220	223	238	243	249	251	254	241	254	232
RESE	113	126	123	117	110	83	101	105	128	119	104	106
BURK	159	169	180	175	164	148	140	135	135	132	136	137
PASA	208	205	218	224	225	233	230	229	227	214	211	215
AZUS	237	237	234	238	242	243	242	240	246	246	248	257
GLEN	233	227	235	233	233	235	241	249	264	288	338	28
POMA	270	273	268	270	265	271	275	279	287	289	292	294
PICO	220	221	226	226	230	231	227	227	222	213	209	197
LYNN	241	245	252	258	259	258	259	260	257	254	247	242
NEWL	164	172	166	175	181	190	196	199	148	126	108	112
LANC	177	186	181	188	190	181	199	202	202	195	192	190
LAHB	245	246	238	246	244	244	242	245	238	237	237	245
ANAH	227	229	229	226	226	223	224	224	219	213	211	202
LSAL	248	244	249	250	257	266	275	278	283	281	283	275
COST	211	219	224	226	230	238	237	232	238	231	224	230
TORO	261	257	250	255	264	254	253	246	236	209	160	145
NORC	281	279	283	285	289	289	294	293	300	299	281	255
RIVR	287	284	281	285	279	282	277	281	286	292	301	314
PERI	348	341	324	322	317	321	330	3	55	92	131	54
HEME	245	239	236	232	234	232	232	223	222	204	193	187
BANN	219	231	236	243	243	242	244	241	248	257	265	263
PLSP	146	111	170	91	315	329	325	319	322	321	324	324
UPLA	249	255	253	251	256	255	254	257	264	271	285	299
FONT	241	244	254	257	255	255	248	245	247	244	258	265
SNBO	219	220	216	220	225	222	216	212	210	212	208	201
RDLA	285	285	285	275	271	267	271	278	294	298	358	93
CRES	200	198	207	223	218	204	194	202	224	232	242	244

Table 3. Average wind direction from 31 SCAQMD Telemetry System Sites for the period May - September 1988 at all 24 hours. Data is in units of degrees; the direction is that from which the wind is blowing.

HOUR STATION	0	1	2	3	4	5	6	7	8	9	10	11
CELA	80	91	86	85	92	93	95	89	73	66	54	54
WSLA	93	91	89	77	85	87	89	86	75	59	44	38
HAWT	77	81	85	87	84	79	91	81	76	72	51	38
LGBH	96	101	102	102	96	94	95	86	78	74	63	54
WHIT	98	93	88	85	78	78	82	88	90	80	76	66
RESE	70	75	78	85	87	91	97	94	86	74	77	75
BURK	61	65	69	73	74	71	69	74	78	67	61	63
PASA	79	85	84	84	76	90	94	79	74	74	68	65
AZUS	62	70	77	78	85	79	84	99	74	62	50	45
GLEN	92	85	82	69	65	60	66	85	69	57	47	45
POMA	61	62	72	80	89	84	94	80	67	69	59	48
PICO	55	67	70	77	81	90	100	89	76	61	46	49
LYNN	64	68	73	78	79	77	79	78	71	64	61	45
NEWL	77	76	77	67	70	74	73	86	84	91	79	66
LANC	55	72	67	74	66	67	80	90	82	83	88	88
LAHB	95	98	95	98	101	94	102	95	87	71	59	54
ANAH	68	75	77	78	79	82	83	76	70	66	55	47
LSAL	88	96	99	96	98	96	94	95	89	79	73	66
COST	91	95	94	88	94	97	90	94	101	91	77	74
TORO	80	80	82	77	81	80	74	80	82	77	67	72
NORC	76	74	76	78	79	77	76	87	88	78	68	66
RIVR	75	76	89	74	90	86	89	84	82	80	67	62
PERI	101	99	96	101	97	93	94	82	89	86	88	90
HEME	61	62	74	73	84	77	89	82	63	64	58	58
BANN	32	34	38	43	38	43	54	60	70	72	71	68
PLSP	34	42	38	44	41	49	52	79	82	87	91	93
UPLA	61	66	64	68	63	60	72	81	66	63	50	44
FONT	77	72	79	70	70	68	73	93	69	54	50	49
SNBO	78	91	84	88	85	92	81	86	80	75	64	60
RDLD	84	79	75	70	64	64	67	92	80	66	50	53
CRES	58	57	63	60	61	58	70	74	55	57	59	59
HOUR STATION	12	13	14	15	16	17	18	19	20	21	22	23
CELA	40	35	29	30	25	23	29	33	44	54	66	73
WSLA	32	30	29	28	29	31	36	52	68	85	90	99
HAWT	23	19	19	17	20	16	22	25	39	40	51	68
LGBH	51	53	52	50	49	55	39	42	51	65	81	90
WHIT	62	48	44	42	47	51	52	53	57	67	76	91
RESE	70	67	66	77	94	93	104	100	88	83	79	75
BURK	57	60	59	60	56	62	59	51	59	61	52	57
PASA	54	48	41	36	40	34	31	43	51	60	68	72
AZUS	43	33	34	30	27	25	22	27	45	54	59	65
GLEN	48	41	41	37	34	32	37	60	74	82	90	92
POMA	48	40	40	33	33	30	32	33	35	46	49	59
PICO	42	39	30	28	30	29	27	29	29	35	42	44
LYNN	39	32	29	27	22	27	27	25	29	34	45	55
NEWL	60	52	53	51	56	62	78	88	97	89	77	77
LANC	87	75	65	64	63	60	58	49	47	50	50	49
LAHB	48	38	32	35	35	38	37	40	46	62	71	88
ANAH	39	33	28	29	32	32	34	37	39	49	57	64
LSAL	66	64	63	64	64	62	63	64	64	73	80	86
COST	71	68	67	67	66	63	64	66	70	74	80	81
TORO	65	57	57	58	61	55	59	62	74	87	96	89
NORC	49	40	36	30	30	30	26	29	44	58	68	76
RIVR	46	40	35	34	30	28	32	39	49	53	56	72
PERI	82	72	61	57	57	60	68	88	88	97	98	89
HEME	59	58	60	48	43	41	34	32	42	45	42	51
BANN	66	68	66	57	50	48	46	36	31	30	29	30
PLSP	101	93	102	99	102	95	83	54	35	25	32	35
UPLA	35	35	33	29	30	30	29	27	37	45	48	60
FONT	39	40	34	32	26	24	25	26	41	61	67	81
SNBO	51	47	49	38	36	35	28	34	42	45	64	75
RDLD	47	45	38	39	35	33	34	47	73	92	88	91
CRES	54	58	57	58	54	57	66	69	66	65	69	66

Table 4. Standard deviation of the wind direction from 31 SCAQMD Telemetry System Sites for the period May - September 1988 at all 24 hours. Data is in units of degrees.

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

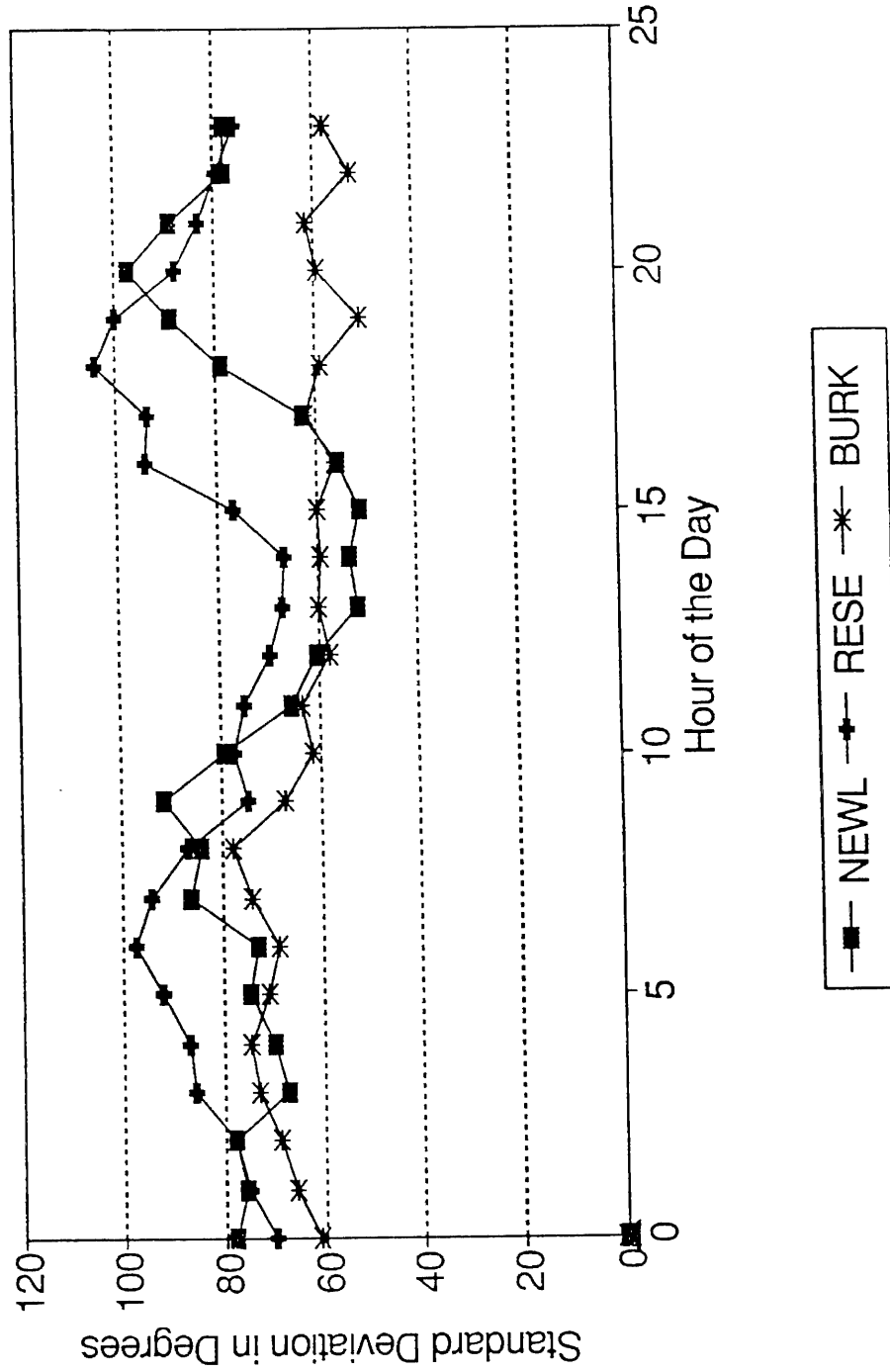


Figure 1. Northwestern Basin Sites in the San Fernando Valley

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

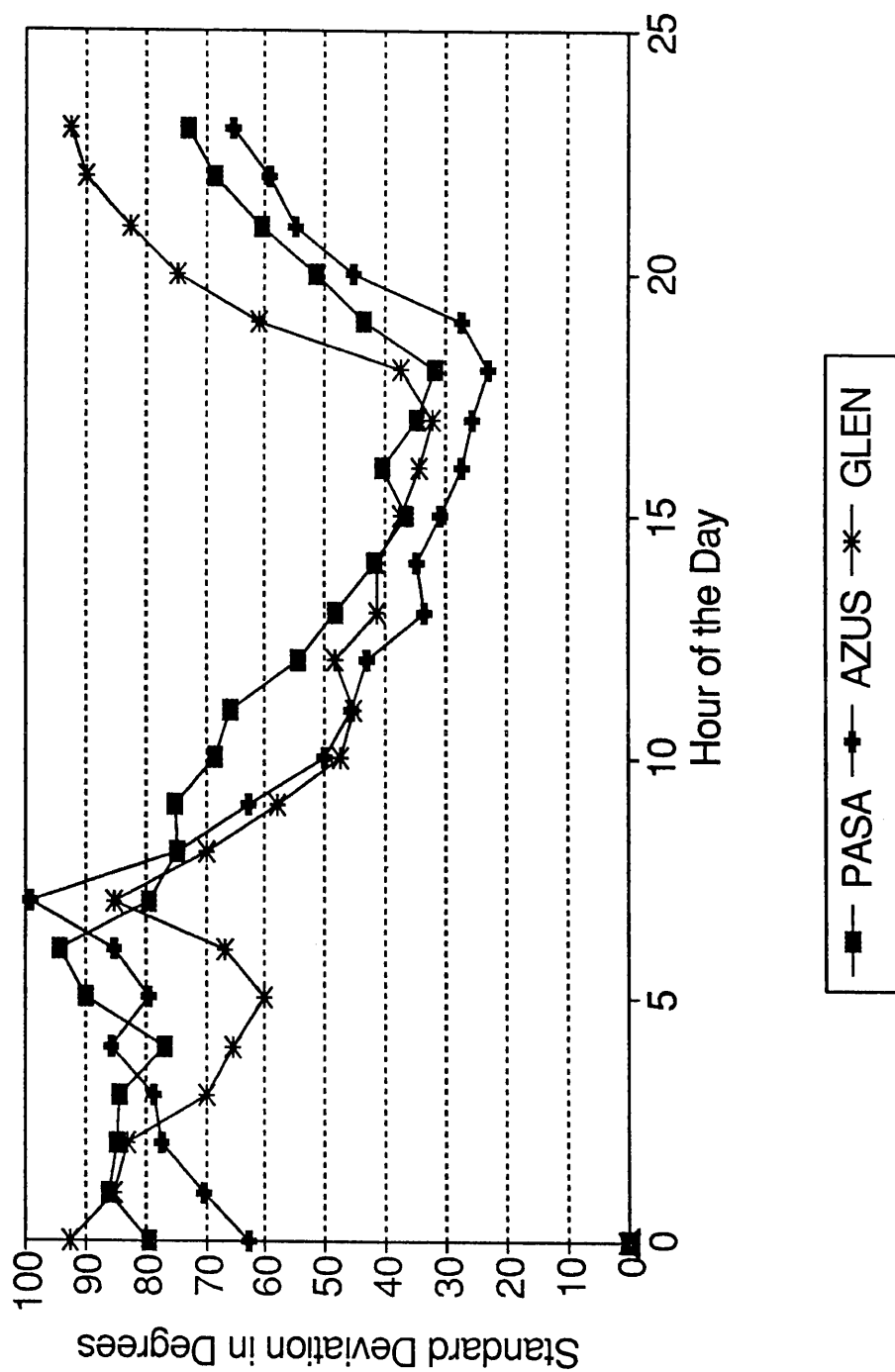


Figure 2: Northern Basin Sites Bordering Mountains

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

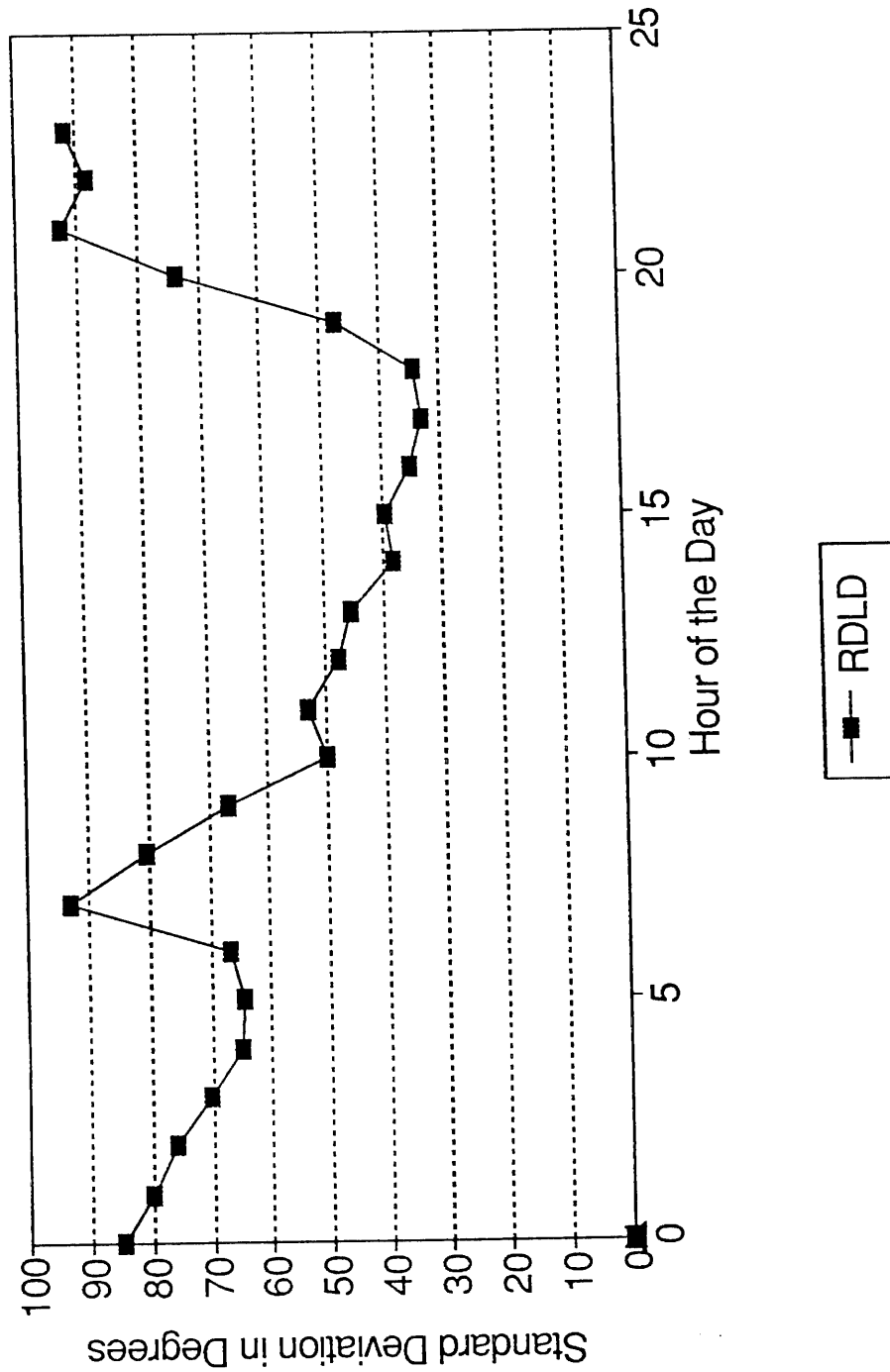


Figure 3. North Central Site Bordering Mountains with a Strong Nighttime Slope Flow

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

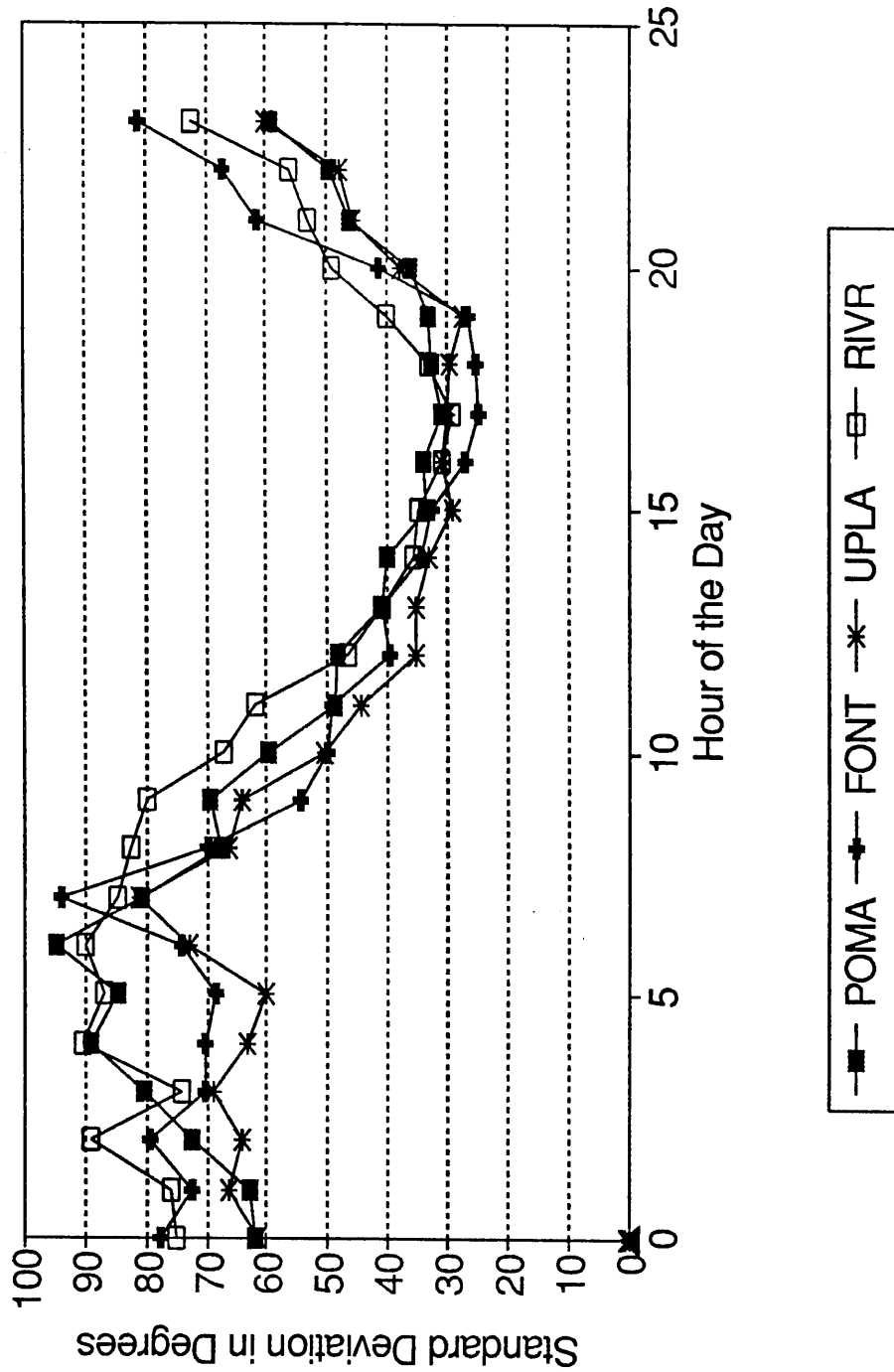


Figure 4. Eastern Basin Sites (1)

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

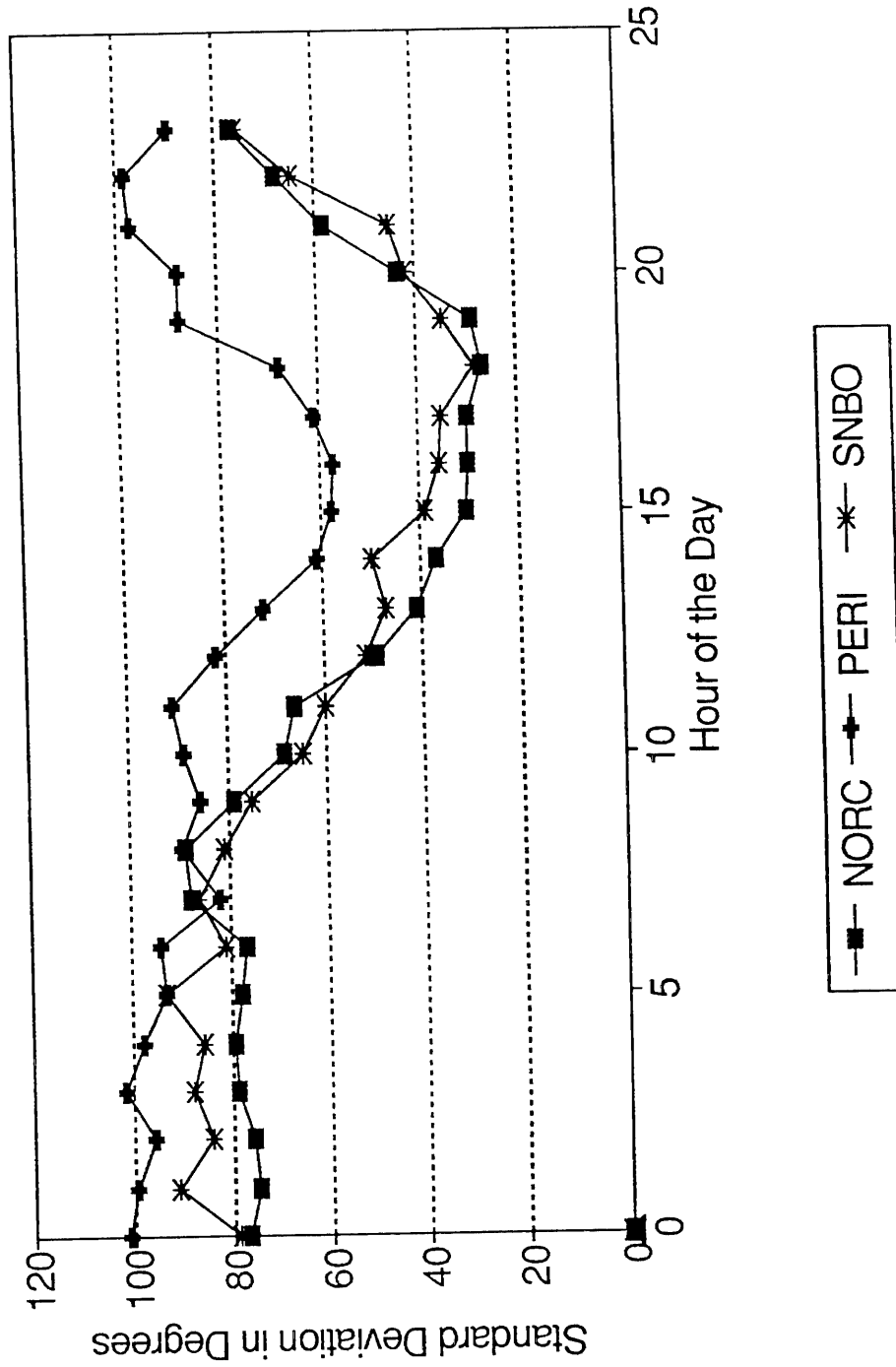


Figure 5. Eastern Basin Sites (2)

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

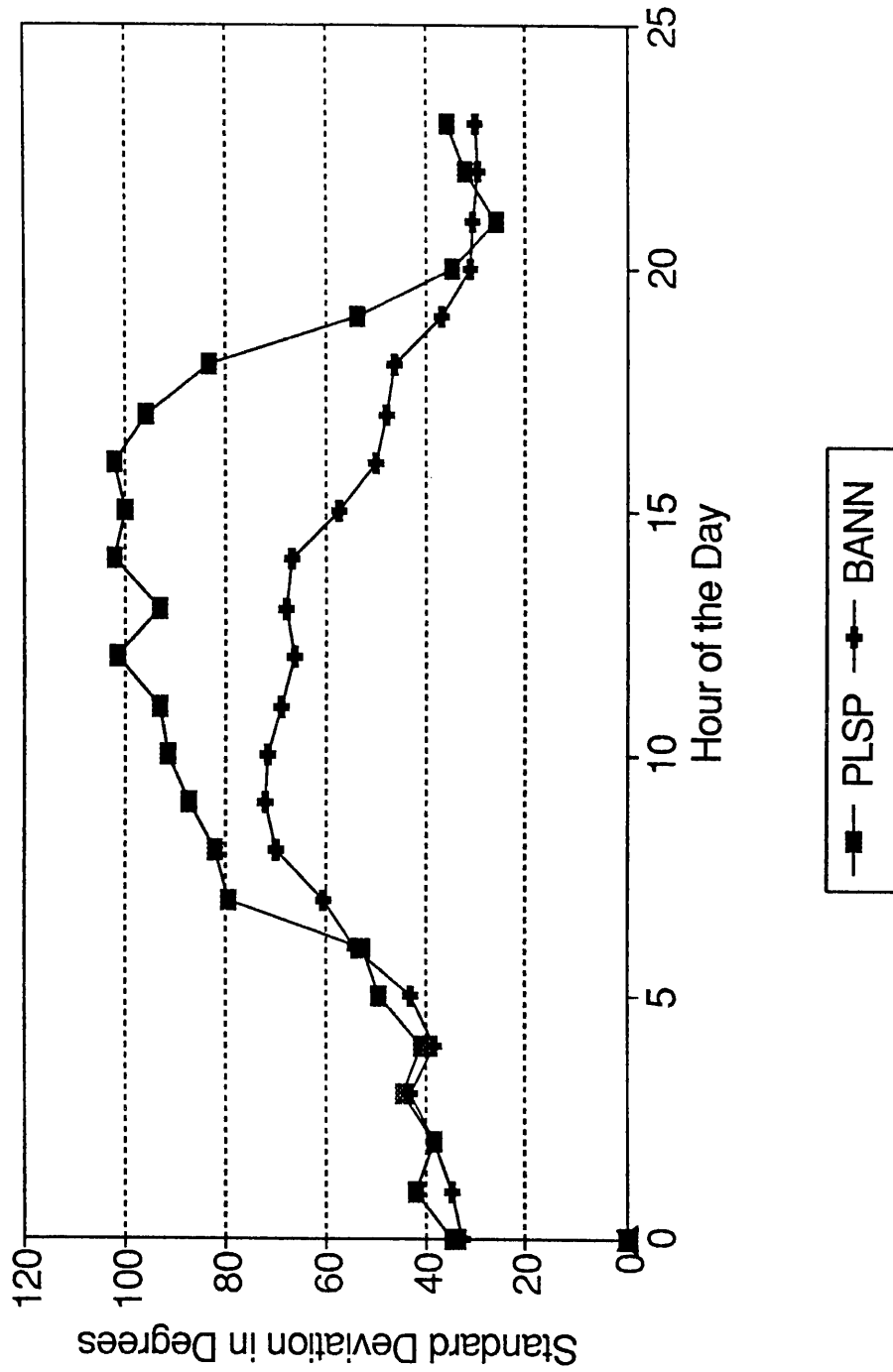


Figure 6. Eastern Desert Sites

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

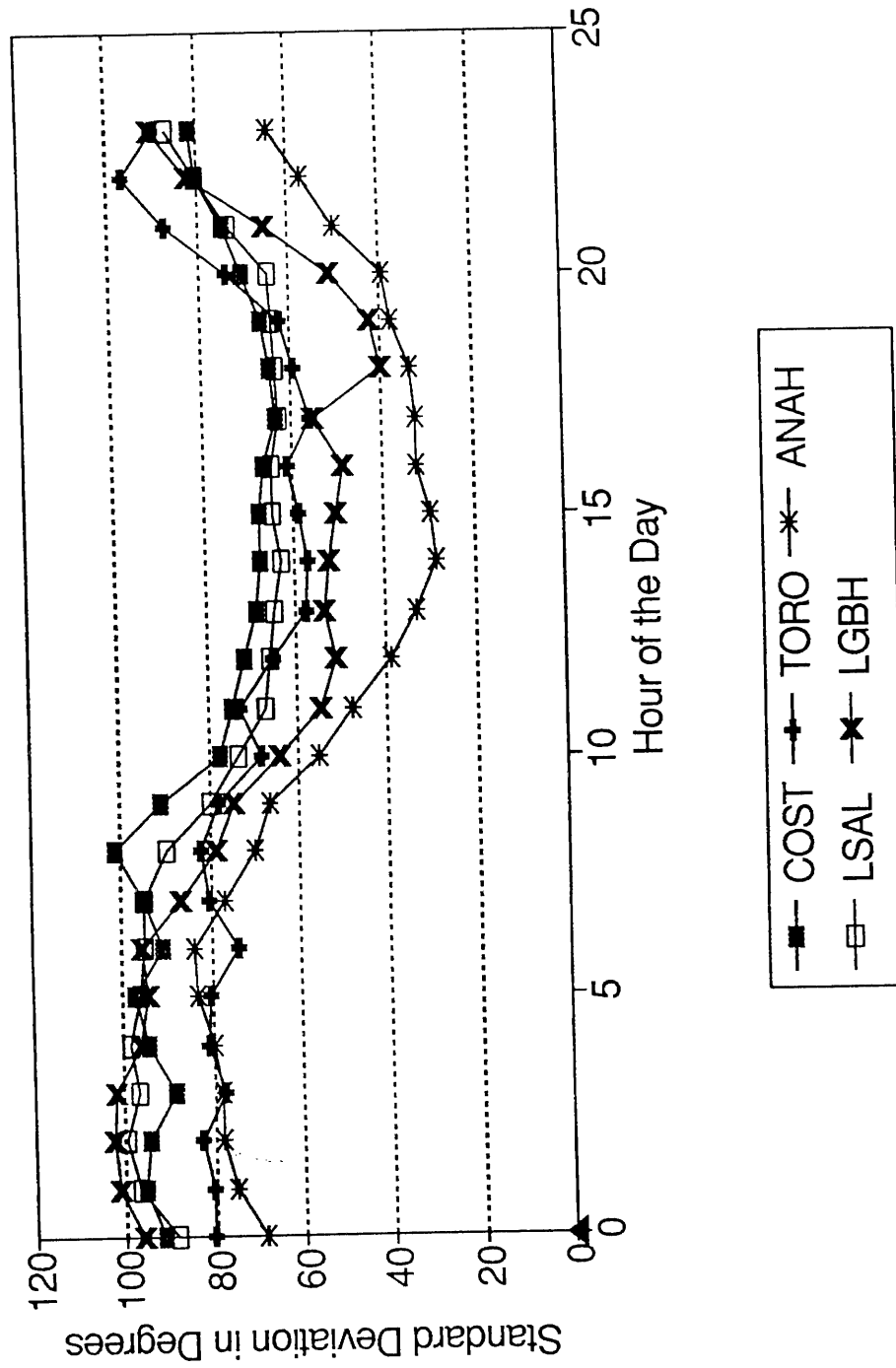


Figure 7. Southwestern Coastal Regions

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

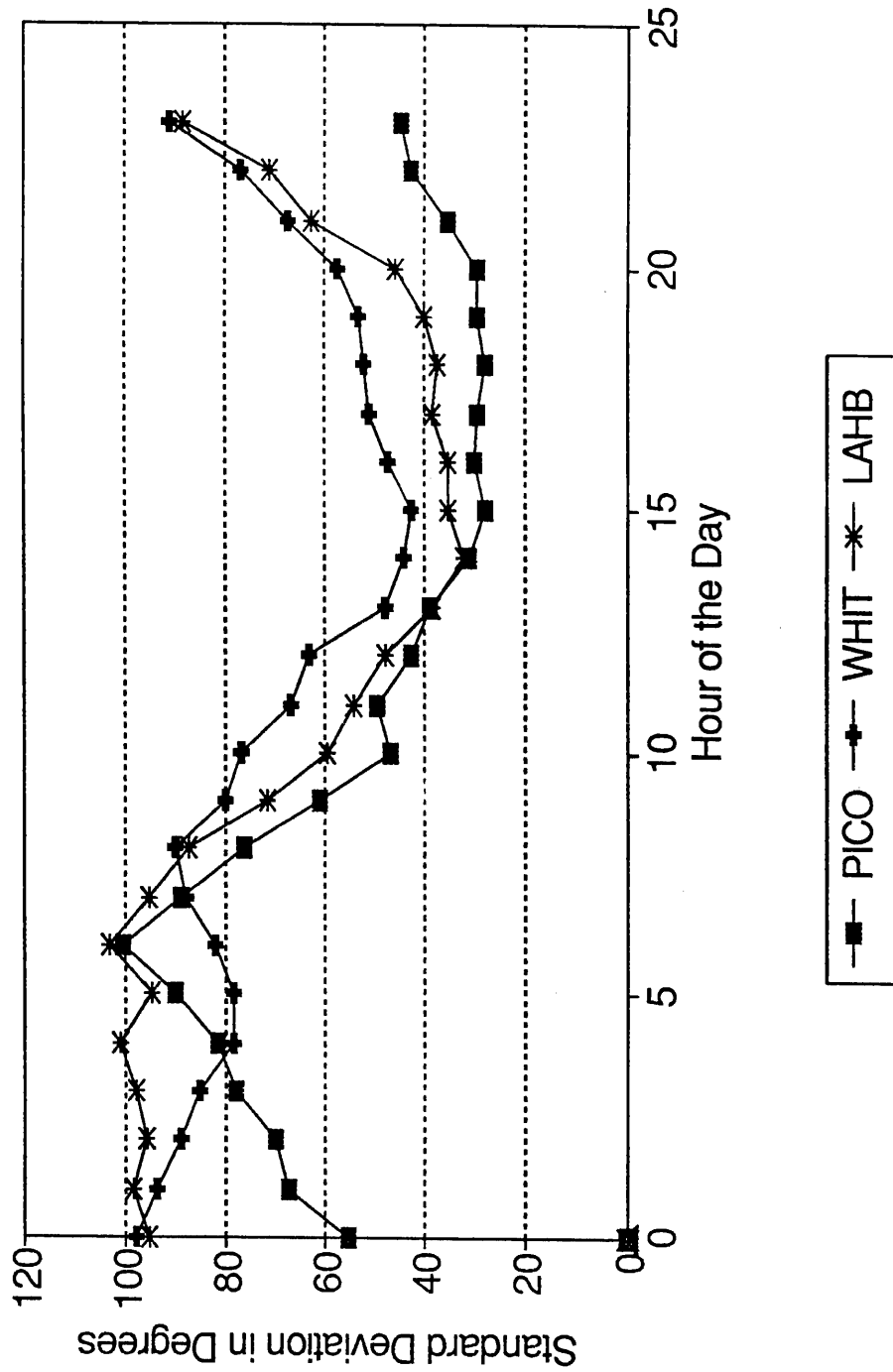


Figure 8. Central Basin Sites

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

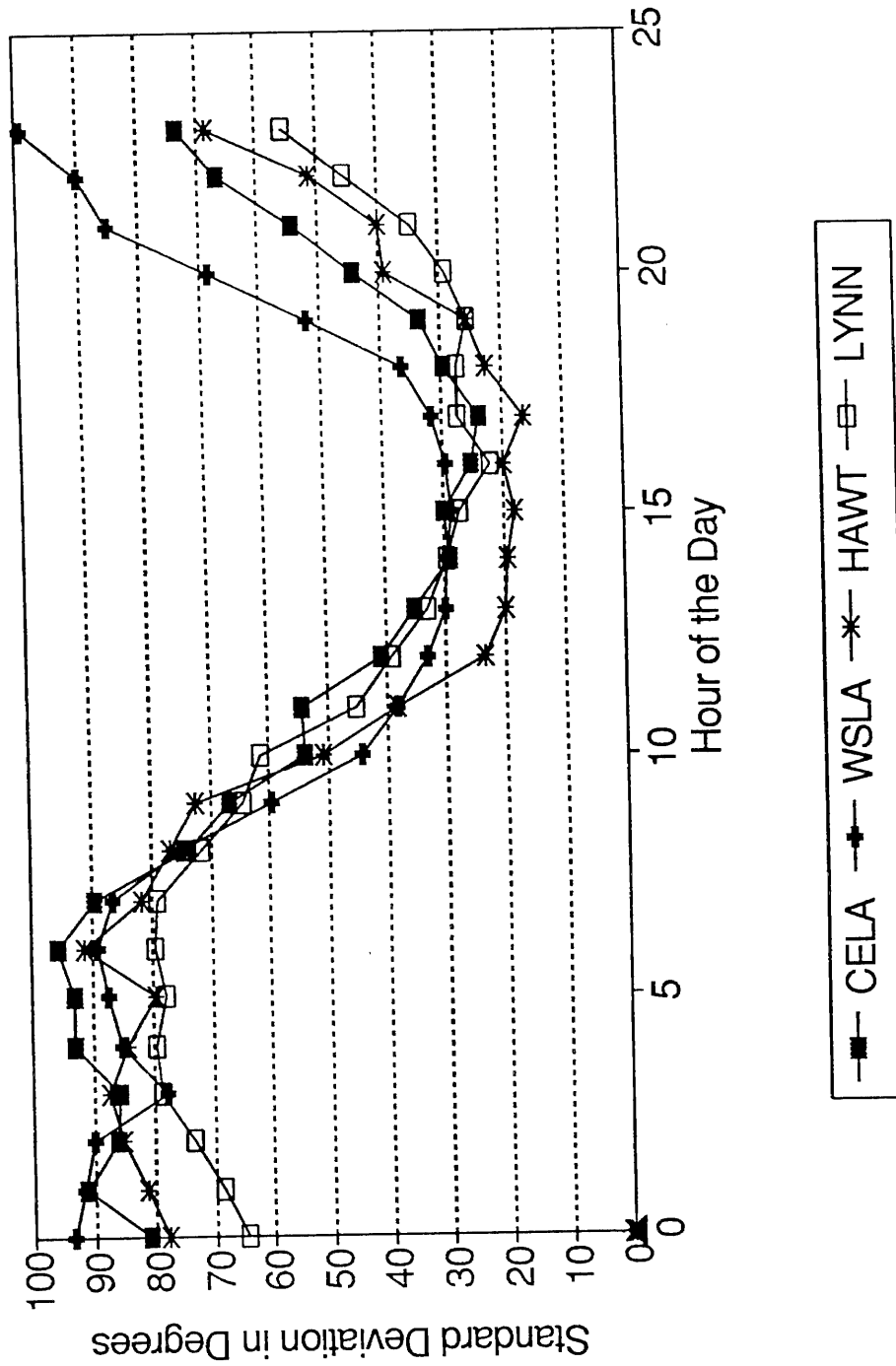


Figure 9. Downtown and Western Coastline Sites

Standard Deviation of Wind Direction

LA Basin Sites May - September 1988

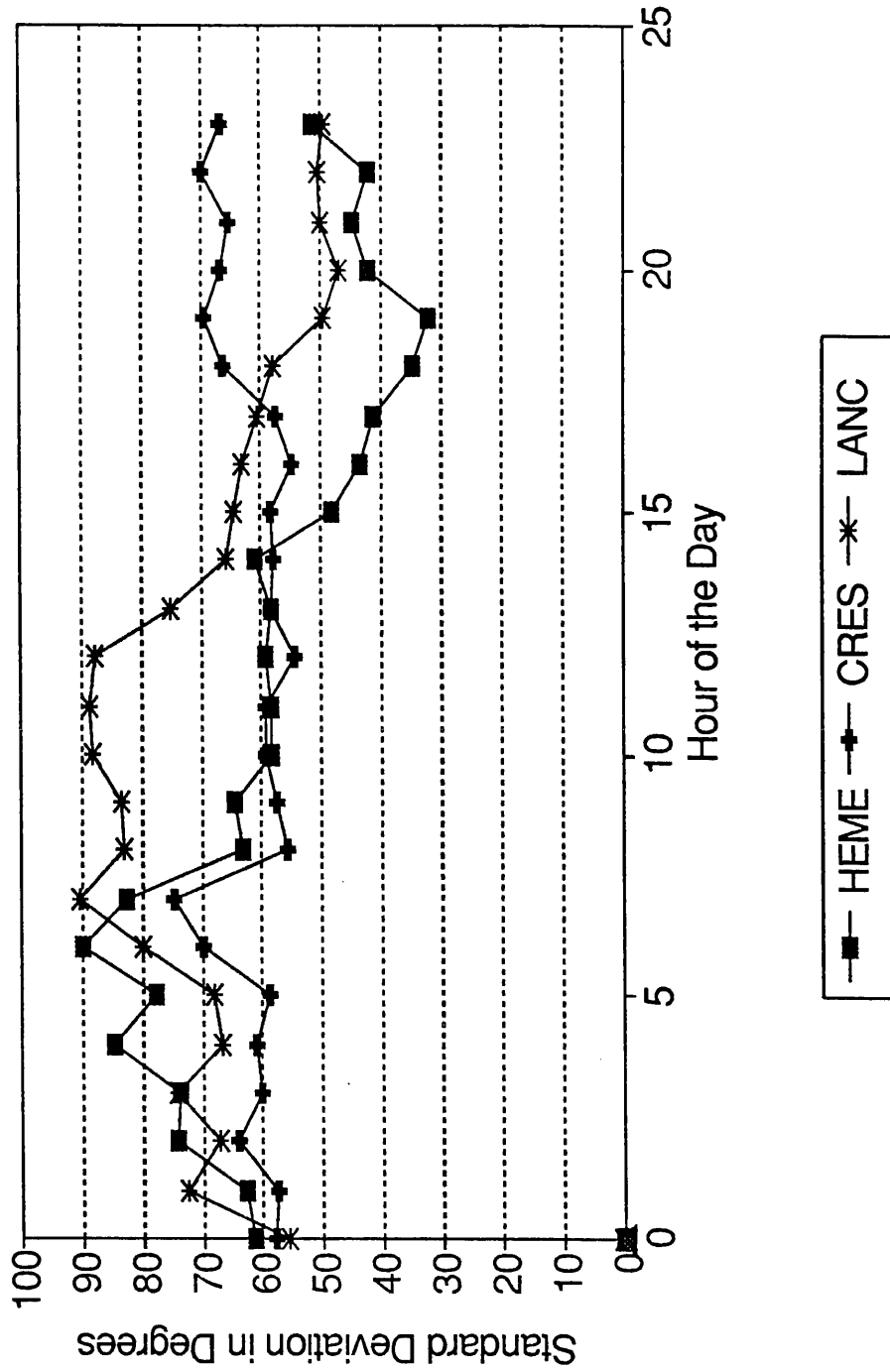


Figure 10. Sites on the Basin's Perimeter

References

1. Arritt R. W. and Pielke A. R. (1986), Interactions of nocturnal slope flows with ambient winds, *Boundary-Layer Meteorology*, **37** no. 1/2, pp. 183-195.
2. Ball F. K. (1956), The theory of strong katabatic winds, *Aust. J. Phys.*, **9**, pp. 373-386.
3. Banta R. M. (1982), Thesis: An observational and numerical study of mountain boundary-layer flow, Department of Atmospheric Science, Colorado State University, paper no. 350.
4. Buettner K. J. K. and Thyer N. (1966), Valley winds in the Mt. Reinier area, *Arch. Met. Geoph. Biok.*, **14**, pp. 125.
5. Burmeister L. C. (1983), Convective Heat Transfer, Chapter 12, John Wiley and Sons, pp. 511-575.
6. Chiba O. and Kobayashi S. (1986), A study of the structure of low-level katabatic winds at Mizuho Station, East Antarctica, *Boundary-Layer Meteorology*, **37** no. 1/2, pp. 343-355.
7. Chico T. et al. (1990), Assessing urban airshed model performance in the South Coast air basin using tracer gases, *Air Quality Mangagement Draft Report*.
8. Defant F. (1951), Local winds. *Compendium of Meteorology*. American Met. Soc. Boston, pp. 655-672.
9. Doran J. C. and Horst T.W. (1982-3), Observations and models of simple nocturnal slope flows, *J. Atmos. Sci.*, **40**, pp. 708-717.
10. Ellison T.H. and Turner J.S. (1959), Turbulent entrainment in stratified flows, *J. Fluid Mech.*, **6**, pp. 423-448.
11. Fitzjarrald D. R. (1984), Katabatic wind in opposing flow, *J. Atmos. Sci.*, **41**, pp. 1143-1158.
12. Gosink J. P. (1987), Entrainment by katabatic winds in Adelie Land, Antarctica, Preprints Third International Symposium on Stratified Flows, **2**.
13. Gutman L. N. (1983), On the theory of the katabatic slope wind, *Tellus*, **35A** no 3, pp. 213-218.
14. Horrell R. S. (1987), Atmospheric buoyancy driven downslope flows with entrainment from a ground source, Masters Thesis, Department of Chemical Engineering, California Institute of Technology, Pasadena, CA 91125.
15. Huppert E., Stephen R., Sparks J., Wilson J. R. and Hallworth M. A. (1986), Cooling and crystallization at an inclined plane, *Earth and Planetary Science Letters*, **79**, pp. 319-328

16. Kondo J. and Sato T. (1988), A simple model of drainage flow on a slope, *Boundary-Layer Meteorology*, **43**, pp. 103-123.
17. Leslie L. M. and Smith R.K. (1974), A numerical study of katabatic winds and their effect on pollutant dispersal in urban areas, *Australasian Conf. on Hydraul and Fluid Mech*, 5th, Proc, Univ. of Canterbury, Christchurch, NZ pp. 553-560.
18. Mahrt L. (1982), Momentum Balance of Gravity Flows, *American Met. Soc.*, **39**, pp. 2701-2711.
19. Manins P. C. and Sawford B. L. (1979), Katabatic winds: A field case study, *Quart J. R. Met. Soc.*, **105**, pp. 1011-1025.
20. Manins P. C. and Sawford B. L. (1979), A model of katabatic winds, *Journal of Atmospheric Sciences*, **36**, pp. 619-630.
21. Ohata T. and Higuchi K. (1979), Gravity Wind on a Snow Patch, *Journal of the Meteorological Society of Japan*, **36**, pp. 254-263.
22. Pielke R. A. (1984), *Mesoscale Meteorological Modeling*, Academic Press, Inc., pp. 23-41.
23. Prandtl L. (1952), *Essentials of Fluid Dynamics*, Blackie and Son Limited, pp. 452. (Also available in its original form in German, 1942.)
24. Rao K. S. and Snodgrass H. F. (1981), A nonstationary nocturnal drainage flow model, *Boundary-Layer Meteorology*, **20**, pp. 309-320.
25. Tennekes H. and Lumley J. L. (1972), *A First Course in Turbulence*, The MIT Press, pp. 27-58.

Chapter Two

A Study of Atmospheric Transport and Dispersion Near a Two-Dimensional Divergence Zone

Section 2.1

Published and presented as part of the Air and Waste
Management Association's 82nd Annual Meeting,
June, 1989.

Submitted for Publication in Atmospheric Environment.

89-138.2
Ground Release SF₆ Tracer Experiments Used to Characterize
Transport and Dispersion of Atmospheric Pollutants
During the Southern California Air Quality Study of 1987

Robin Scott Horrell, Michael Deem, Peter Wyckoff, and Fred Shair
Chemical Engineering, California Institute of Technology
Pasadena, CA 91125

and

Neil Crawford
Chemical Engineering, University of California, Berkeley
Berkeley, CA 94704

March 25, 1989

ABSTRACT

As part of the 1987 South Coast Air Quality Study (SCAQS) program, 4 SF₆ tracer tests were conducted. The tracer results help document the transport and dispersion of the air over downtown Los Angeles. An objective trajectory algorithm was developed in order to use both surface and upper level winds to help interpret the tracer data.

Based upon the spread of the tracer during the summer, the effective horizontal eddy diffusivity was estimated to be of order 10^4 meters²/second. Negligible "carry over" was observed during the summer.

The extent of "carry over" during the winter was estimated from mass balances of the tracer. During the winter studies, the residence time of the tracer within the Los Angeles Basin exceeded 36 hours.

The relationship of the tracer to observations of elevated pollutant concentrations are examined.

INTRODUCTION

Los Angeles has had one of the most comprehensive pollution control programs in the country. Despite this effort many of the inhabitants of the LA Basin are exposed to ozone concentrations which exceed the federal standard more than 1 out of every 3 days, Davidson¹, (1986). By carefully considering the effects of meteorology, Davidson has shown that the control program has lead to a steady decline in ozone exposure. Yet the rate of decline he reports over the 1976-1984 period, if extrapolated, is not enough to bring Los Angeles into compliance with the Clean Air Act until beyond the year 2000. In addition to ozone, other airborne pollutants such as carbon monoxide, nitrogen dioxide, and PM_{10} often exceed both state and federal standards. Clearly measures which go beyond those already in effect need to be devised, studied, and then implemented in order to improve the basin's air quality.

Relatively speaking, a great deal is known about the transport and dispersion that takes place during the daily on-shore flow meteorological regime. Much less is known about the characteristic of the **nighttime off-shore flow** meteorological regime. Still less is known about the **morning and evening transition regimes** which often coincide with the so-called "rush hours". Almost nothing is known about the impact of emissions which are **transported out to sea** during the night and then returned during the subsequent on-shore flow. Likewise, almost nothing is known about pollutants that are transported into the surrounding mountains during the daytime **up-slope flow** and returned during the subsequent nighttime **down-slope flow**. Finally, almost nothing is has been documented regarding the influence of the winds "aloft" upon the transport and dispersion of pollutants throughout the Los Angeles basin. In short, more needs to be documented and understood about (i) **flows across the boundaries of the region of interest**, (ii) **the transition regimes**, and (iii) **the flows aloft**. This information is essential for the efficient development of new control measures aimed at improving air quality.

Thus, four sulfur hexafluoride (SF_6) tracer studies were conducted in order to document the transport and dispersion of certain ground level emissions during the Southern California Air Quality Study (SCAQS) in 1987. Although not all of the above mentioned needs could be addressed in only 4 tests, the above issues were kept in mind during the planning the tracer tests. The SF_6 tracer releases were coordinated with experiments conducted by Tracer Technologies Inc. and Southern California Edison that involved the release of perfluorocarbons near ground level and aloft, England² et al., (1989).

The tracer studies were conducted (i) to help improve forecast models, and (ii) to provide data to test numerical model simulations which attempt to predict pollution levels throughout the Los Angeles basin. The SF_6 tracer studies were designed to quantitatively document the transport and dispersion of emissions released in downtown Los Angeles during the morning and evening transition

regimes. Two SF₆ studies were conducted during the summer, and two during the fall.

An objective interpolation and extrapolation model, referred to as **WIND**, was developed in order to rapidly utilize the South Coast Air Quality Management's (SCAQMD) wind telemetry data. Currently **WIND** is run on an IBM PS2 personal computer with a high resolution color monitor. **WIND** used both surface and upper level air data to develop trajectories. The tracer data and the wind trajectories were used to help clarify the mechanisms and patterns of atmospheric transport.

The purpose of this report is (i) to present the pertinent data concerning the transport and dispersion of morning and evening emissions from downtown Los Angeles, and (ii) to discuss the implications of these data.

Brief Review of Previous Studies

Various aspects of the local meteorology have long been known. Edinger^{3,4} (1959) described the marine layer and its affect on the meteorology of the basin, and the distribution of photochemical smog (1973). Information on the three-dimensional transport of pollutants, and carry over in the Los Angeles region during a smog episode was reported by Blumenthal⁵ et al. (1977), and Bytnerowicz⁶ et al. (1987).

Shair⁷ et al. were the first to use a gaseous tracer to quantitatively study the transport and dispersion associated with the land breeze/sea breeze circulation system. They found that elevated emissions released during the nighttime off-shore flow could significantly impact the air quality during the subsequent on-shore flow regime. They also found certain quantitative aspects of the circulation pattern to be rather persistent.

Hayes⁸, (1984) has organized the surface winds in Los Angeles into seven categories. During the summer, the winds fall almost exclusively into 4 of the 7 categories. During the nighttime, between 22:00 PDT and 4:00 PDT, 71% of the time the surface winds were either downslope or calm. During the daytime, between 10:00 PDT and 16:00 PDT, the surface winds fall into either the onshore south or sea breeze categories; both these regimes involve an onshore flow from Santa Monica to the Palos Verdes Peninsula. During the winter, the surface winds are distributed more evenly into the 7 categories. For example, at 4:00 PDT the surface winds are likely to be calm (29%), offshore (25%), downslope (17%), and Santa Ana (17%). At 10:00 PST the flows are similarly grouped but the likelihood of a southerly flow has increased to 15%, while that of the other categories has decreased by a few percent. The afternoon winter winds, like those in the summer, fall primarily into two categories, sea breeze (51%) and onshore south (24%). The evening winds are again diverse and nearly like the winds observed at 4:00 PST.

EXPERIMENTAL

The techniques used during the SCAQS SF₆ were similar to those used by previous investigators. For example see Drivas and Shair⁹ (1974), Drivas¹⁰ (1975), Lamb¹¹ (1978), and Reible¹² (1982). Basically, these investigations consist of releasing SF₆ at a constant known rate from a source (or source region) of interest, and (ii) collecting a large number of air samples throughout region and time of interest. The tracer concentrations within the air samples are determined by means of electron-capture gas chromatography. Such studies have elucidated many source/receptor relationships of interest.

Tracer Releases

Sulfur hexafluoride, (having a molecular weight of 146), was released at a constant rate from 6:00 to 9:00 PDT during July 15, September 3, November 11, and from 16:00 to 19:00 PST during December 10, 1987. The first three tests were to probe the morning transition regime during which the morning "rush hour" occurs. The last day was to probe the evening transition, during which the evening "rush hour" occurs. These days were chosen on the basis of the SCAQS short term meteorological forecasts. The hope was to release the tracer on days during which elevated pollution levels within the Los Angeles region would be observed within the following 36 hours. The release site for each test was the Southern California Edison Office in Vernon, California. This site, (longitude: 118 degrees 14 minutes, latitude: 34 degrees) was chosen because (i) it is centered approximately in the middle of the hydrocarbon emissions of the downtown area, and (ii) representatives of SCE permitted us to use their facility.

During the July 15th study, SF₆ was released at a rate of 426 grams per minute leading to a total release of 76.7 kilograms (or 169 pounds). During the September 3rd study, SF₆ was released at a rate of 1151 grams per minute leading to total release of 207.2 kilograms (or 456 pounds). During the November 11th study, SF₆ was released at a rate of 772 grams per minute leading to a total release of 139 kilograms (or 306 pounds). During the December 10th study SF₆ was released at a rate of release 1362 grams per minute leading to a total release of 245 kilograms (or 539 pounds). In order to compare the results of different tracer experiments, the downwind tracer concentrations must be normalized by means of their molar release rates. The molar release rates for tests 1,2,3 and 4 were 175, 473, 317, and 560 gram moles per hour respectively.

Air Sample Collection

Air samples were collected by means of two methods. First, 5-second "grab" samples were collected by passengers along auto traverses. Second, hourly averaged samples were collected by means of automatic samplers located at several "fixed" sites of interest. All samples were collected in 30 cc plastic syringes that had been previously vacuum cleaned and tested. The auto traverse data were used to recreate a "snap shot" of the tracer cloud.

During Test 1, 30 automatic samplers were deployed at 20 sites. Tables I, II, and III list the telemetry, upper air, and fixed sampling sites; figures 1 and 2 show their locations. During Test 2, 29 automatic samplers were deployed at the same sites as in Test 1. During Test 3, 12 automatic samplers were deployed at 10 different sites. In Test 4, 22 automatic samplers were deployed at 11 different sites.

Unfortunately during Tests 1 and 2, a divergence zone lay to the east of the release site and most of the tracer was transported northward into the San Fernando Valley. Although good descriptions of the clouds were obtained via auto traverses, few hourly averaged SF_6 tracer data were collected since the fixed sites were chosen in anticipation of atmospheric transport from downtown Los Angeles towards the east. The hourly averaged SF_6 tracer data for Test 3 are listed in Tables IVa and IVb. The hourly averaged SF_6 tracer data for Test 4 are listed in Tables Va, Vb, and Vc. Sampling begins on the hour reported and continues for 1 hour.

Analysis of SF_6

The air samples were analyzed by the means of electron capture gas chromatography. The apparatus and technique are described by Drivas and Shair (1975). Concentrations of SF_6 can be readily detected in concentrations down to 1 part tracer per trillion parts (ppt) of air by volume.

TRANSPORT TRAJECTORIES OBTAINED FROM METEOROLOGICAL DATA

A computer model, **WIND**, was developed to objectively interpolate and extrapolate the surface and upper level winds throughout the SCAQS region of interest. **WIND** was used to generate two-dimensional transport trajectories at various elevations. These trajectories were compared to those inferred by the SF_6 tracer data. For any location within the region of interest, a trajectory may be calculated and displayed on the computer's high resolution color screen. A topographical map of the SCAQS region of interest provides the background on which the transport trajectories are displayed. **WIND** also permits rectangles to be drawn around any zone within the SCAQS region of interest; the volume flux of air may be calculated and displayed around the rectangle. Such diagrams indicate the presence of convergence and divergence zones.

The algorithm for interpolation and extrapolation was based upon the desire to have the wind at any point within the SCAQS region of interest to be calculated from at least two telemetry reporting stations. First, the average distance between a telemetry station and its **two nearest neighbors** is determined for each such station. The average value of the all these distances, R_c , is called the "critical radius".

When the **WIND** program is executed, R_c is first calculated and remains constant throughout the subsequent calculations unless changed manually. The **WIND** program automatically calculates a new value of R_c for each new set of

telemetry stations. The interpolation and extrapolation schemes are based upon the following algorithm. To obtain the velocity vector at a point P_i on the map, a circle of radius R_c is drawn with P_i at its center. An imaginary right circular cone of height R_c is constructed having as its base the circle drawn previously. In this way the center of the cone lies at a height R_c above the point P_i . The surface of the cone represents the weighting function which describes how stations surrounding P_i are used to determine the velocity vector at that point. The weight, i.e. the influence a particular telemetry station has on the local velocity is determined by projecting the point which represents the location of that station vertically onto the cone. The height of the point of intersection with the cone is the amount which that station influences the calculation of the velocity vector. Points that lie outside of the radius R_c are given a weight of zero, while the ones inside have weights ranging from 0 to R_c . A single component direction of the velocity vector at a given point is then the normalized sum of all the corresponding single component direction vectors interior to R_c multiplied by their respective weighting factors. The normalization constant is the sum of the weighting factors of all the stations. Thus, this construction is geometrical in nature and, in accordance with intuition, does not attribute influence to telemetry stations far from the point in question. This method is extremely quick because of the minimal computational requirements, and thus is ideal for the PC.

Trajectories are determined by the successive application of the following recursive algorithm at each hourly interval. A point P_1 is specified on the map which represents the center of the "parcel" of air being tracked. The local velocity at that point is calculated in the manner described above. The location of the parcel of air an hour later, P_2 , is estimated by calculating the vector displacement which would occur if the parcel moved at the local velocity at P_1 for an hour. The estimate is assessed and improved according to the following method. The line segment formed by P_1 and P_2 is bisected at P_3 . The velocity at P_3 is determined, and the parcel is allowed to travel at this velocity for half the time of the preceding interval, in this case one half hour, ending up at position P_4 . If the absolute difference between P_2 and P_4 is less than a tolerance (which is inversely proportional to the level of the recursion, in this case $1/2$) then P_2 is accepted as the new location of the air parcel after one hour. However, if the absolute difference between P_2 and P_4 is greater than the tolerance, then the process is repeated at a time interval equal to one half of the previous time interval. That is, the line segment formed by P_1 and P_3 is then bisected and the sequence of steps described above is repeated on this smaller length interval. This process is continued until the tolerance is satisfied on a sub-segment of the total segment P_1P_2 . The recursion then backs its way out satisfying the tolerance requirement at each interval, progressively calculating the intermediate position of the air parcel as it moves during one hour.

The use of the recursion algorithm permits the development of trajectories which are very sensitive to regions where convergence, divergence and abrupt directional changes occur. The increased sensitivity results from assessing the

velocity at more than just one point along the route of an hour long trajectory.

Currently **WIND** is used to examine the wind fields within the Los Angeles region. The region of coverage spans from 117 degrees 7 minutes to 118 degrees 17 minutes in latitude, and from 33 degrees 15 minutes to 34 degrees 25 minutes in longitude. The region corresponds to that bounded at the north west corner by Newhall, at the north east corner near Crestline, at the south east corner near Perris and at the south west corner in the Pacific Ocean. The inputs for **WIND** are the data obtained from the SCAQMD's telemetry system along with the upper level wind data gathered during the SCAQS program. The stations within the telemetry system are listed in Table I and depicted in Figure 1. The upper level wind stations are listed in Table II, and depicted in Figure 2 (Lehrman¹⁴, 1988).

RESULTS

July 15, 1987

About 12.5 cubic meters (at normal conditions) of pure SF₆ were released from 6:00 - 9:00 PDT, (the morning transition period) near ground level from the Vernon site. The extent to which the tracer dispersed during the morning transition period was determined from data collected via auto traverses. As shown in Figure 3, by 9:30 PDT the tracer cloud had dispersed over a region with horizontal dimensions 22 kilometers in the north/south direction by 17 kilometers in the east/west direction. Although not apparent in Figure 3, the tracer cloud was almost circular - with an approximate radius of 11 kilometers (7 miles). By considering the solution to the transient radial diffusion equation, the "effective eddy diffusivity" was estimated to be of the order of 10⁴ meters²/second.

The mixing depth over the downtown area at 9:30 PDT was estimated to be about 500 meters (Bennett¹³, 1988; Lehrman¹⁴, 1988). Thus, by 9:30 PDT, the tracer had mixed through a volume of approximately equal to 1.8×10^{11} cubic meters. The average concentration of the tracer within this volume was calculated to be 80 ppt. Consequently, the tracer cloud was estimated to contain 14.4 cubic meters of pure SF₆. Thus, **taking into account the inherent uncertainties in the mass balance calculation**, it is reasonable to conclude that the tracer cloud contained essentially all of the tracer that was released.

The mid day surface and upper level winds swept the tracer northward and then westward into the San Fernando Valley. The **WIND** generated trajectories are shown in Figure 5a. As indicated, the morning southerly winds shifted into the typical onshore flow. Upper level wind trajectories which start at 6:00 PDT from Vernon, and continue for 18 hours, are depicted in Figure 5c. The surface wind trajectories lagged behind the transport of the tracer. The observation that the surface winds underestimate the transport distance of a tracer is in agreement with the observations of Neiburger¹⁵, 1955. However, the faster upper level wind trajectories are in excellent agreement with the tracer data shown in Figures 2 and

3. Between 100 and 400 meters AGL over Vernon, the upper level winds were observed to rotate counter clockwise with increasing altitude, (see Figure 5c).

September 3, 1987

About 34.7 cubic meters, (at normal conditions) were released from 6:00 - 9:00 PDT near ground level at the Vernon site. As shown in Figures 4a and 4b, the extent to which the tracer had dispersed at various times were again determined from auto traverse data. As shown in Figure 4a, rather high concentrations of tracer were observed up to 7 miles south of Vernon between 9:00 - 10:00 PDT. It should be noted that neither the available surface winds nor the available upper level winds gave any indication of transport to the south of the release site.

As in the previous test, the mid day winds swept the tracer northward and then westward into the San Fernando Valley. As in the previous test, the winds between 100 and 400 meters AGL over Vernon were found to rotate counter clockwise with increasing altitude, (see Figure 5d). The winds aloft were stronger than those in the previous test. Note that in both the summer tests, the morning transition period was characterized by light, but directional, surface winds.

The **WIND** generated trajectories indicate that the upper level winds are required to explain the location of the tracer cloud that was observed during the afternoon. Trajectories of the surface winds were found to lie to the east of the tracer in the afternoon. The 100 and 200 meter AGL trajectories were found to coincide with the location of the tracer. The 300 and 400 meter AGL trajectories were found to lie to the west of the tracer. As in the first test upper level wind trajectories were required to explain the location of the tracer.

Ozone Concentrations during Tests 1 and 2

The concentrations and distributions of ozone during Tests 1 and 2 were remarkably similar. For example in both tests, maximum hourly averaged ozone concentrations occurred between 13:00 and 15:00 PDT. These times coincided with the approximate arrival times of the tracer. The maximum ozone concentrations at Burbank, Reseda, and Newhall were 12-14 pphm, 13-17 pphm, and 17-18 pphm respectively. Thus, the tracer data indicate that the atmospheric pollutants, residing north and west of the divergence zone, contributed to the observed ozone in the San Fernando Valley. During the morning, the divergence zone existed along a strip which passed south and east of Vernon, and extended to Pasadena.

Elevated ozone concentrations were also observed in the eastern part of the Los Angeles Basin. For example the maximum hourly averaged ozone concentrations at Fontana, San Bernardino, Redlands, and Crestline were 16-18 pphm, 18-19 pphm, 13-17 pphm, and 20-22 pphm respectively. Trajectories shown in Figures 5a and 5b indicate that atmospheric pollutants to the south and east of the divergence zone - and north of Costa Mesa, contributed to the ozone in the eastern region of the Los

Angeles Basin. The trajectories starting at Costa Mesa indicate that the transport of atmospheric pollutants south of Costa Mesa was influenced by the Elsinor convergence zone.

By 14:00 PDT the tracer cloud had reached the dimensions of 43 kilometers x 27.5 kilometers (see Figure 4b). The "effective eddy diffusivity" associated with this cloud was calculated to be of the order of 10^4 meters²/second - in excellent agreement with that calculated from the previous test. The mixing depth was estimated to be 300 meters (Bennett¹³, 1988; Lehrman¹⁴, 1988). Thus, during the afternoon, the tracer was dispersed within a volume of about 3.5×10^{11} cubic meters within which the average concentration of tracer was 108 ppt. Consequently, the amount of tracer contained within the cloud was estimated to be 37.8 cubic meters of pure SF₆ - indicating that the tracer cloud shown in Figure 4b accounted for essentially all of the tracer that was released.

Also shown in Figure 5b are the trajectories for sites located in Orange County. It is interesting to note that the southern most site shows trajectories that exhibit an abrupt change in direction in the region of the Elsinor convergence zone.

November 12-13, 1987

About 23.3 cubic meters of pure SF₆ were released from 6:00 - 9:00 PDT. The hourly averaged tracer data are listed in Table 4a. The northern most station at which SF₆ was observed was Burbank. The low concentrations seen at Burbank (from 13:00 - 15:00 PDT) may have been transported by the surface level winds depicted in Figure 8a. During the release, the tracer was transported to the west and apparently offshore. Following the flow reversal which transported the tracer onshore, the West Los Angeles site experienced an hourly averaged concentration of 136 ppt. Furthermore, significant tracer concentrations were observed as far south as Lynwood (from 9:00 - 14:00 PDT). As indicated by the trajectory in Figure 9a, the tracer at Lynwood may have been transported by the 100 meter AGL winds. These observations indicate the great extent to which the tracer spread as a result of a flow reversal. It should be noted that SF₆ was observed at sites within the basin up to 36 hours following the start of the release.

Auto traverse data collected from 6:00 - 8:00 PDT on November 13 indicated the presence of a large cloud of tracer (with dimensions 37 km x 37 km) as indicated in Figure 6a. The average concentration of tracer within this region was 21 ppt. Assuming a mixing depth of 200 meters would account for about 25% of the tracer released during the morning of the previous day. By then, the tracer had undergone at two flow reversals. The amount of tracer lost due to vertical mixing during these flow reversals is unknown.

It should be noted that 36 hour surface wind trajectories, and 18 hour upper level wind trajectories, do not leave the SCAQS region of interest. The computed

trajectories, shown in Figure 8a indicate that much of the tracer may have been transported into the San Gabriel Mountains. However, the location of the maximum concentration of SF_6 , (observed between 6:00 - 8:00 PDT on November 13), was 35-45 kilometers to the east of the corresponding trajectories.

December 10-11, 1987

Approximately 41.1 cubic meters of pure SF_6 were released from 16:00 - 19:00 PST (during the evening transition period) on December 10. The transport of the tracer, offshore and then onshore, can be inferred from the hourly averaged data listed in Tables Va, Vb, and Vc. Tracer concentrations above 1000 ppt were seen in central Los Angeles for 7 hours following the end of the release. From 1:00 - 6:00 PST on December 11, concentrations above 100 ppt were seen moving eastward over West Hollywood; between 2:00 and 3:00 PST the average concentration was 311 ppt. Eleven hours later this cloud apparently was transported along the foothills of the San Gabriel mountains; the average SF_6 concentration from 13:00 - 14:00 was 290 ppt. The surface wind trajectories, in qualitative agreement with above mentioned observations, indicated a clockwise transport of a portion of the tracer. However, the trajectories do not show indicate the fact that the majority of the tracer was transported over the ocean - where there were no meteorological data available. Following the release, the westward transport of the tracer is shown in Figure 6b.

As shown in Figures 7a and 7b, the majority of the tracer was seen moving onshore south of Vernon during the morning and afternoon of December 11. The average concentrations of tracer within the dashed boxes of Figures 7a and 7b are 54 ppt and 46 ppt respectively. Assuming a mixing depth of 200 meters indicates that during the morning of December 11, 87% of the tracer was contained in the dashed box (with dimensions of 61 km x 53 km) shown in Figure 7a. Even in the afternoon, about 21% of the tracer was observed over the western region of the Los Angeles Basin (see Figure 7b); the dimensions of the box are 24 km x 36 km. During the morning and early afternoon, the surface winds along the coast were on shore. These observations indicate that a considerable portion of the SF_6 returned to the Los Angeles region after having resided offshore overnight. The extent of the offshore flow 19 hours earlier can be inferred from the fact that SF_6 was coming onshore during the **afternoon of December 11**. Offshore meteorological data would be needed to predict the degree to which the tracer was dispersed south of the release site.

Some of the tracer was transported offshore around the Palos Verdes Peninsula, and then onshore as far south as Corona Del. Emissions, from a single low level point source, were transported and dispersed over an area of 3000 square kilometers surrounding the release site.

Elevated concentrations of carbon monoxide were observed in the south west portion of the basin. Values near the 20 ppm California state standard,

(for a 1 hour average concentration) were observed Lynwood from 6:00 - 8:00 PDT on December 11. The corresponding concentration of tracer was 45 ppt.

CONCLUSIONS

- (1) In both summer studies, during the hour following the end of a three hour release, the tracer was dispersed over an area of approximated 400 square kilometers.
- (2) During the summer, an effective eddy diffusivity of around 10. meters²/second characterized the dispersion of the tracer.
- (3) Mass balances of the tracer indicated that essentially all of the tracer was accounted for all of the tracer.
- (4) An objective trajectory model was developed to help interpret the tracer data.
- (5) Both the surface and upper level winds are needed to explain the transport of the tracer.
- (6) The morning downtown emissions did not impact the ozone in the eastern region of the Los Angeles region.
- (7) In both winter studies, a vast majority of the tracer was carried over. The residence time within the SCAQS area of interest was greater than 36 hours.
- (8) Mass balances indicated that a more complex pattern of dispersion occurred during the winter as compared to that during the summer.
- (9) The convoluted surface wind trajectories help verify the presence of flow reversals.
- (10) In both cases the tracer was transported outside of the boundary for which adequate meteorological data are available.
- (11) Increased meteorological data, included (i) winds over the ocean, (ii) winds in the mountains, and (iii) upper levels winds are needed to adequately reconstruct all of the transport and dispersion patterns.
- (12) During December 11, 1987, the tracer was observed to coincide with the elevated carbon monoxide levels at Lynwood.

Acknowledgements:

This work was funded in part by a grant from the California Air Resources Board. We are delighted to acknowledge the excellent support given to us by (i) Charles Bennett, Frank DiGenova, Eric Fugita, and Manjit Ahuja of the California Air Resources Board in Sacramento, and (ii) Howard Bakes, Don Daymon, Pat Harrington, Lee Lewis, Loudes Menes, Janet Nguyen, Frank Ong, and King Yu of the California Air Resources Board in El Monte. We also wish to thank William Bope and Joseph Cassmassi of the South Coast Air Quality Management District for their valuable help and insight. Neil Crawford was supported by an NSF Research Experience for Undergraduate (REU) grant administered through the Caltech Summer Undergraduate Research Fellowship (SURF) program. We wish to thank Carolyn Merkel, Joan Spears, and Jean Cass of the SURF office.

REFERENCES

1. A. Davidson, "Comment on ten-year ozone trends in California and Texas," *APCA Journal*, **36** (5): 597 (1986).
2. W. G. England, L. H. Teuscher, S. Marsh, "Perfluorocarbons Tracer Experiments during the Southern California Air Quality Study," *Air and Waste Management Association*, paper 89-138.1 (1989).
3. J. G. Edinger, "Changes in the depth of the marine layer over the Los Angeles Basin," *J. Met.* **16** 3 219 (1959).
4. J. G. Edinger, "Vertical distribution of photochemical smog in Los Angeles Basin," *Env. Sci and Tech.* **7** 3 247 (1973).
5. D. L. Blumenthal et al., "Anatomy of a Los Angeles smog episode: pollutant transport in the daytime sea breeze regime," *Atmos. Env.* **12** 893 (1977).
6. A. Bytnerowicz et al., "Gaseous and particulate air pollution in the San Gabriel Mountains of Southern California," *Atmos. Env.* **21** 8 1805 (1987).
7. F. H. Shair et al., "Transport and dispersion of airborne pollutants associated with the land breeze-sea breeze system," *Atmos. Env.* **16** (9) 2043 (1982).
8. T. Hayes et al., *California Surface Wind Climatology*, State of California Air Resources Board Aerometric Data Division, 12 (1984).
9. P. J. Drivas and F. H. Shair, "A tracer study of pollutant transport and dispersion in the Los Angeles area," *Atmos. Env.* **8** 1155 (1974).
10. P. J. Drivas, *Investigation of Atmospheric Dispersion Problems by Means of a Tracer Technique*, Doctoral Thesis California Institute of Technology, Pasadena CA 91125 (1975).
11. B. K. Lamb, *Development and Application of Dual Atmospheric Tracer Techniques for the Characterization of Pollutant Transport and Dispersion*, Doctoral Thesis California Institute of Technology, Pasadena CA 91125 (1978).
12. D. D. Reible, *Investigations of Transport in Complex Atmospheric Flow Systems*, Doctoral Thesis California Institute of Technology, Pasadena CA 91125 (1982).
13. C. Bennett, Personal communication, California Air Resources Board, Sacramento CA 95812 (1988).
14. D. Lehrman, Personal communication, Technical and Business Systems, Santa Rosa, CA 95404 (1988).
15. M. Neuburger, "Tracer tests of the accuracy of trajectories computed from the observed winds in the Los Angeles area," *Air Pollution Foundation*, (1955).

Table I Station Code and Name for SCAQMD Telemetry Stations

Code	Station Name	Code	Station Name
ANAH	ANAHEIM	AZUS	AZUSA
BANN	BANNING	BURK	BURBANK
CELA	LOS ANGELES	COST	COSTA MESA
CRES	CRESTLINE	FONT	FONTANA
GLEN	GLENDORA	HAWT	HAWTHORN
HEME	HEMET	LAHB	LA HABRA
LANC	LANCASTER	LGBH	LONG BEACH
LSAL	LOS ALAMITOS	LYNN	LYNWOOD
NEWL	NEWHALL	NORC	NORCO
PASA	PASADENA	PERI	PERRIS
PICO	PICO RIVERA	PLSP	PALM SPRINGS
POMA	POMONA	RDLD	REDLANDS
RESE	RESEDA	RIVR	RIVERSIDE
SNBO	SAN BERNARDINO	TORO	EL TORO
UPLA	UPLAND	WSLA	WEST LOS ANGELES
WHIT	WHITTIER		

Table II Station Code and Name for SCAQS Upper Air Stations

Code	Station Name
BURB	BURBANK
EL	EL MONTE
LONG	LONG BEACH
LOYO	LOYOLA MARYMOUNT UNTV.
ONTA	ONTARIO
RIVE	RIVERSIDE

Table III Code and Name for Additional Automatic Sampler Locations

Code	Station Name
ALTA	ALTADENA
CORO	CORONA DEL MAR
LADE	LADERA HEIGHTS
PALO	PALOS VERDES PENINSULA
RUBI	RUBIDOUX
WHLL	WEST HOLLYWOOD

Tables 1, 2, and 3.

Table IVa Automatic Sampler Results, November 12-13
November 12, 1987

TIME	6	7	8	9	10	11	12	1	2	3	4	5	PST
SITE													
HAWT	0	0	0	405	382	48	30	14	11	5	6	4	
WSLA	0	0	0	0	9	58	136	77	17	2	0	0	
GLEN	0	0	0	0	0	0	0	0	4	6	9	83	
ELMO	0	0	0	0	0	0	0	50	33	18	16	28	
BURK	0	0	0	0	0	0	0	18	20	0	5	5	
ANAH	0	0	0	0	0	0	0	12	18	2	4	5	
PICO	0	0	0	0	0	0	35	11	0	37	7	4	
LYNN	0	0	0	36	24	367	8	44	19	2	4	5	
POMA	0	0	0	0	0	0	0	0	0	0	0	47	
PASA	0	0	0	0	0	0	0	0	47	34	25	11	

Table IVb
November 13, 1987

TIME	6	7	8	9	10	11	12	1	2	3	4	5	PST
SITE													
HAWT	4	6	9	10	14	6	9	6	12	14	11	15	
PASA	20	32	28	14	26	21	23	12	6	9	3	7	

Tables 4a and 4b.

Table Va Automatic Sampler Results, December 10-11
December 10, 1987

TIME	3	4	5	6	7	8	9	10	11	12	1	2	PST
SITE													
PASA	0	0	0	0	0	0	0	0	0	0	0	0	
LYNN	0	0	0	84	69	46	27	19	22	31	25	28	
HAWT	0	0	49	38	48	70	100	210	10	19	186	38	
CELA	0	6	79	237	1176	1349	2254	1146	828	1353	1418	67	
LGBH	0	0	0	0	0	0	0	0	0	16	17	20	
WHLL	0	0	0	0	0	0	0	0	0	0	261	311	

Table Vb
December 11, 1987

TIME	3	4	5	6	7	8	9	10	11	12	1	2	PST
SITE													
PASA	0	0	0	0	0	0	0	8	40	34	290	257	
ALTA	0	0	0	0	0	0	0	0	19	13	48	41	
CORO						13	8	12	53	13	9	11	
LADE	61	18	40	29	33	33	49	44	38	36	30	47	
WHLL	205	119	128	138	67	21	32	35	27	33	101	22	
PICO			0	0	0	10	13	7	48	24	28		
RUBI									0	0	0		

Table Vc
December 11, 1987

TIME	3	4	5	6	7	8	9	10	11	12	1	2	PST
SITE													
PASA	270	240	115	79	41	54	23	26	42	17	22	24	
ALTA	19	20	18	10	14	9	13	14	6	9	10	13	
CORO	19	18	23	11	10								
PICO	27	0	0										
RUBI	0	0	0	0	0	0	15	3	15				

Tables 5a, 5b, and 5c.

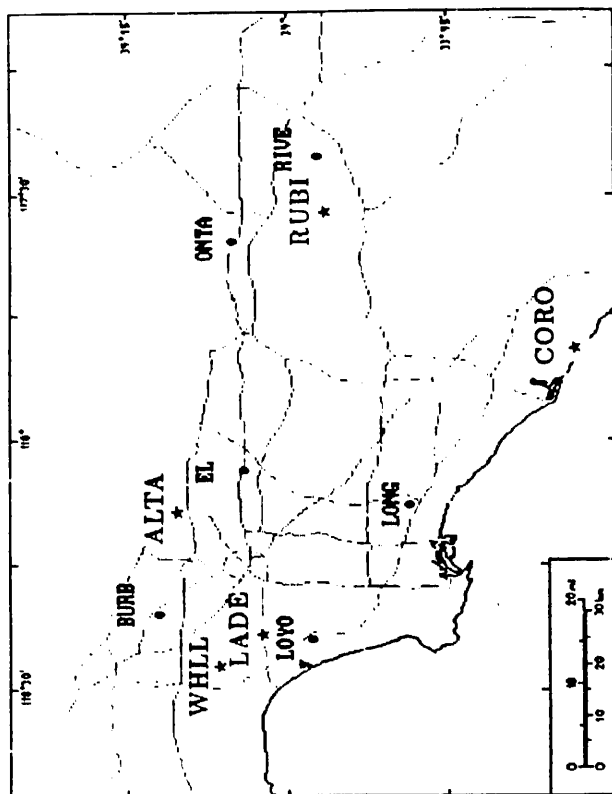
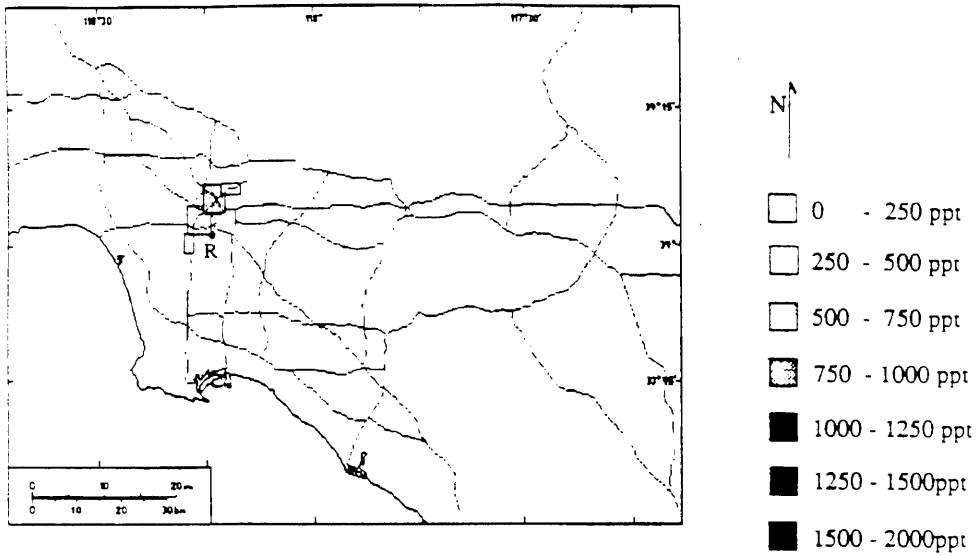


Figure 2. Location of upper air network stations •, and additional one hour average sample board locations *.

(a)



(b)

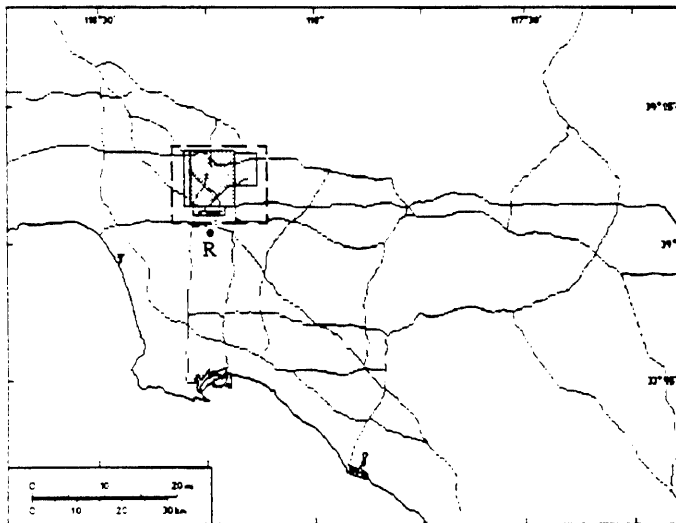
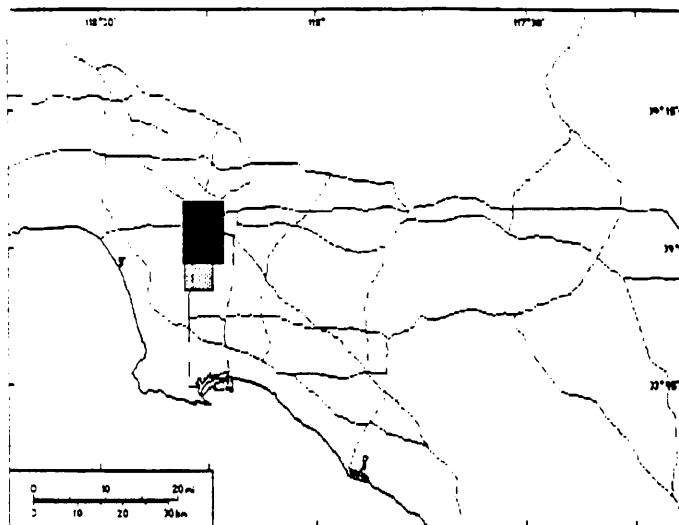
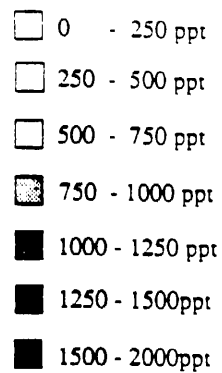


Figure 3. SF₆ tracer cloud as seen by auto traverses on July 15, 1987, (a) 9:00 - 9:30 PDT and (b) 9:30 - 10:30 PDT.

(a)



N



R = Vernon Release Site

(b)

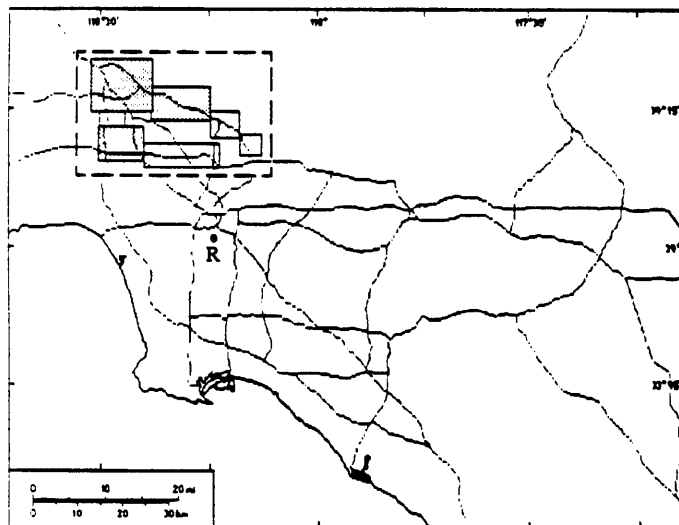


Figure 4. SF_6 tracer cloud as seen by auto traverses on September 3, 1987, (a) 9:00 - 10:00 PDT, and (b) 13:00 - 14:00 PDT.

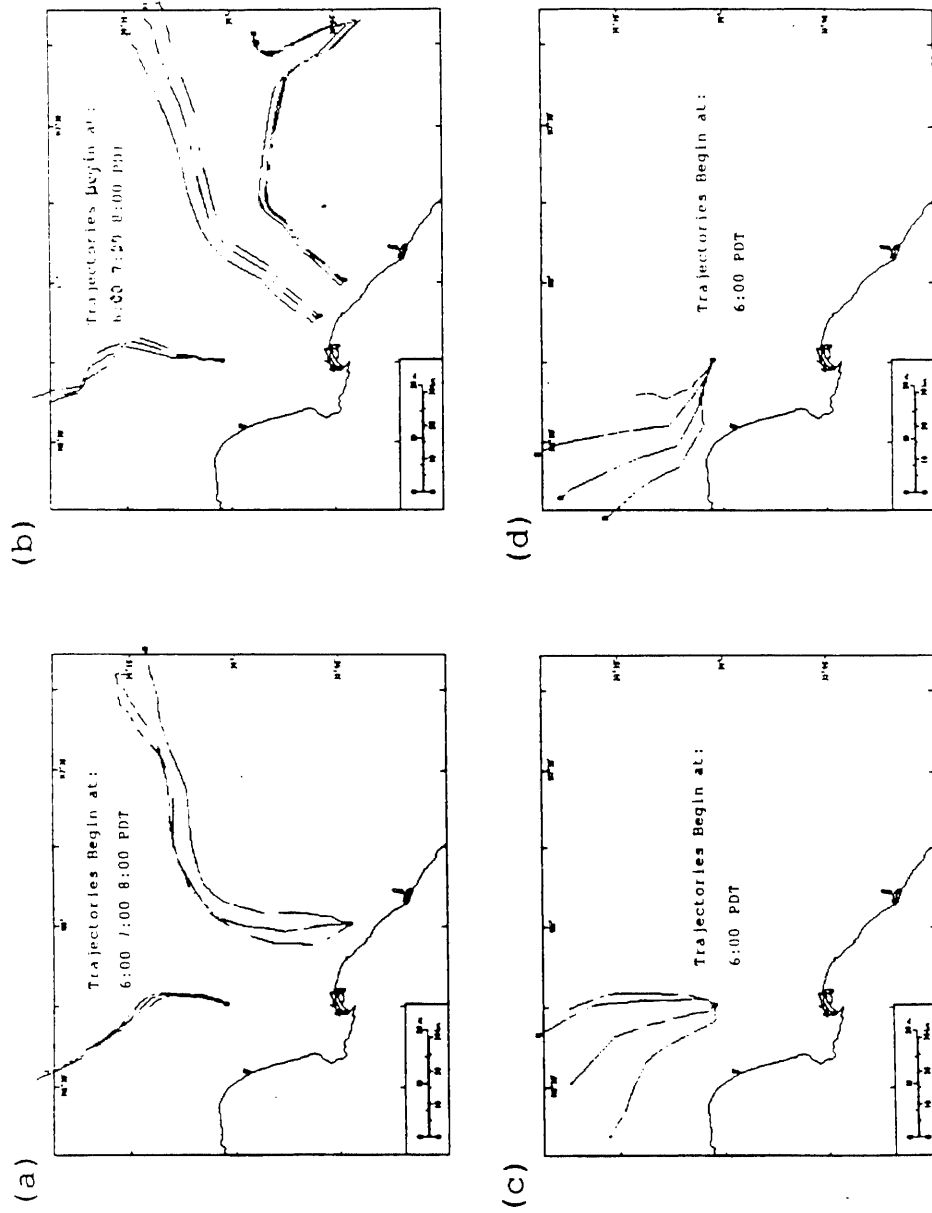
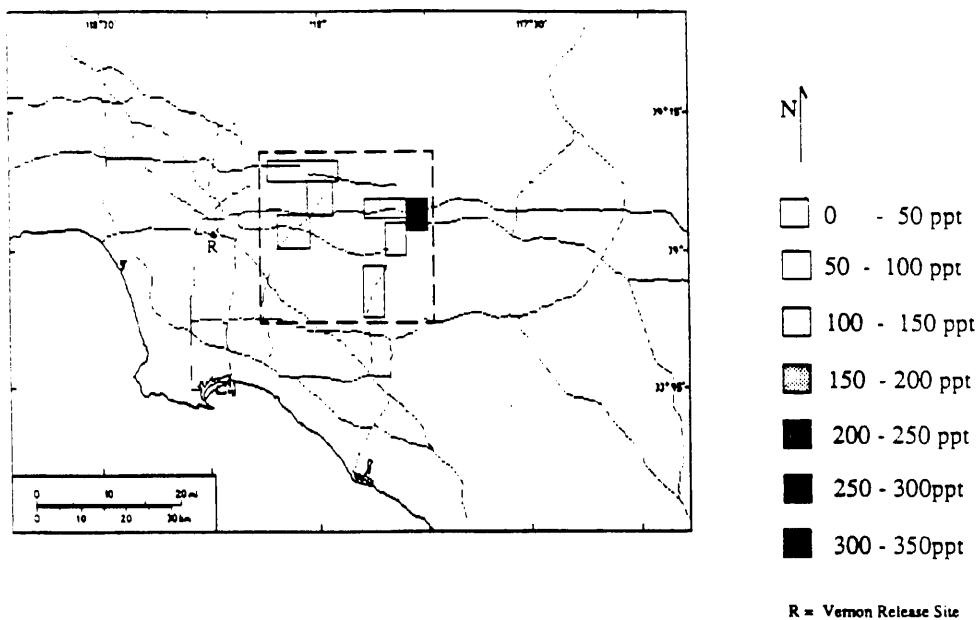


Figure 5. Surface wind trajectories from Vernon and other locations at 6, 7, and 8 PDT, and lasting 24 hours (a) July 15, 1987, (b) September 3, 1987, and upper air trajectories starting at 6 PDT for altitudes of 100, 200, 300, and 400 meters AGL going counter-clockwise about Vernon (c) July 15, 1987, and (d) September 3, 1987.

(a)



(b)

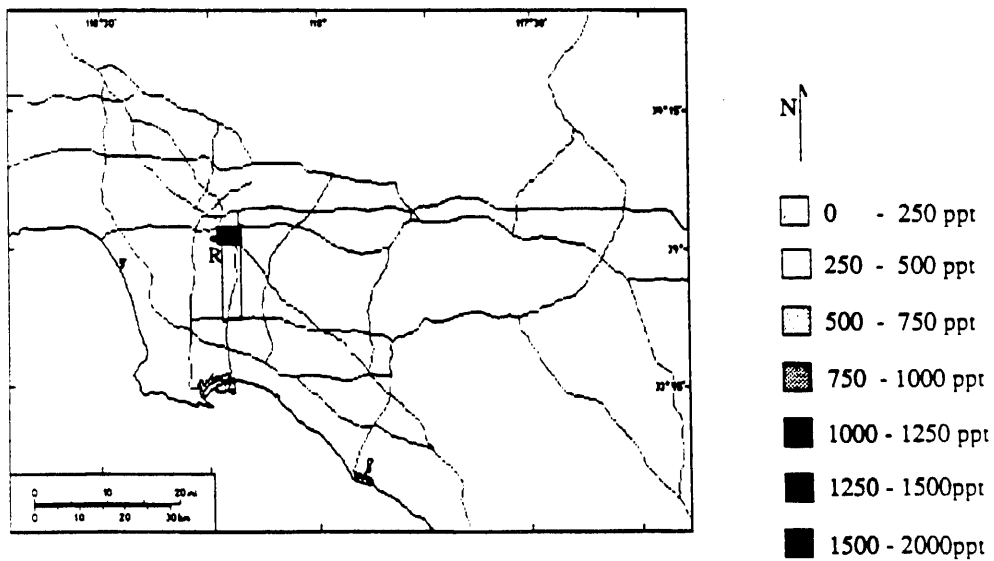
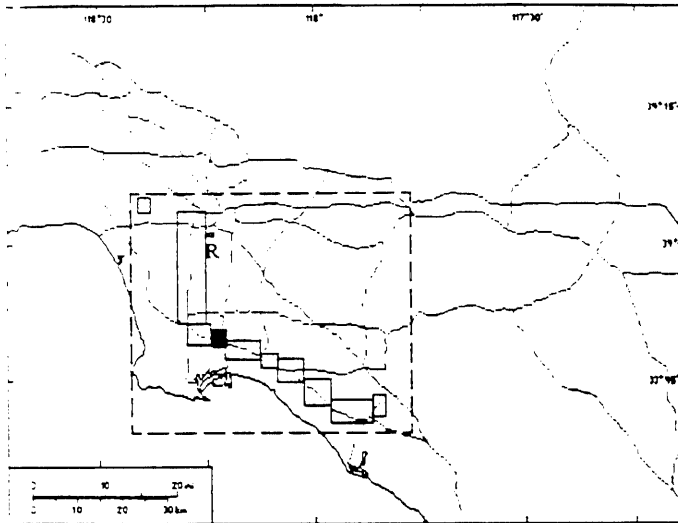
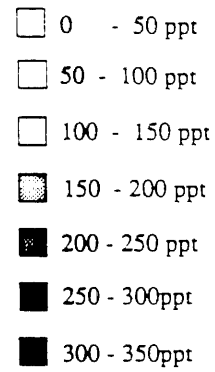


Figure 6. SF_6 tracer cloud as seen by auto traverses on (a) at 6:00 - 8:00 PST November 13, 1987, 24 hours after the release, (b) at 19:00 PST December 10, 1987, just after the evening release.

(a)



N



R = Vernon Release Site

(b)

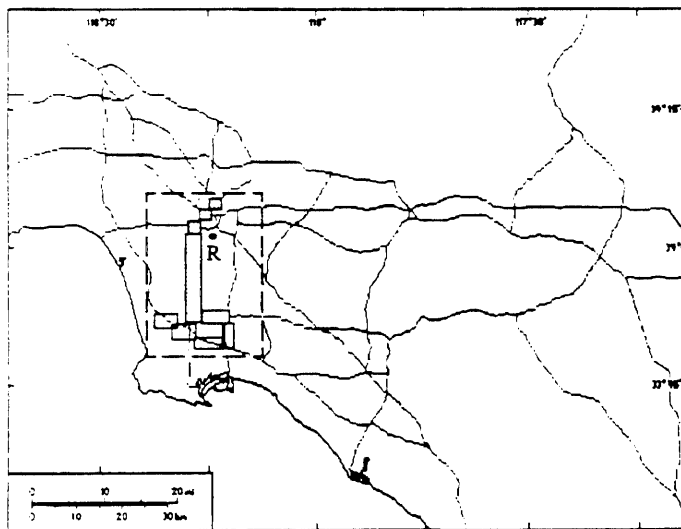


Figure 7. SF_6 tracer cloud as seen by auto traverses on December 11, 1987, the day after the release
 (a) at 7:35 PST, (b) at 14:00 PST.

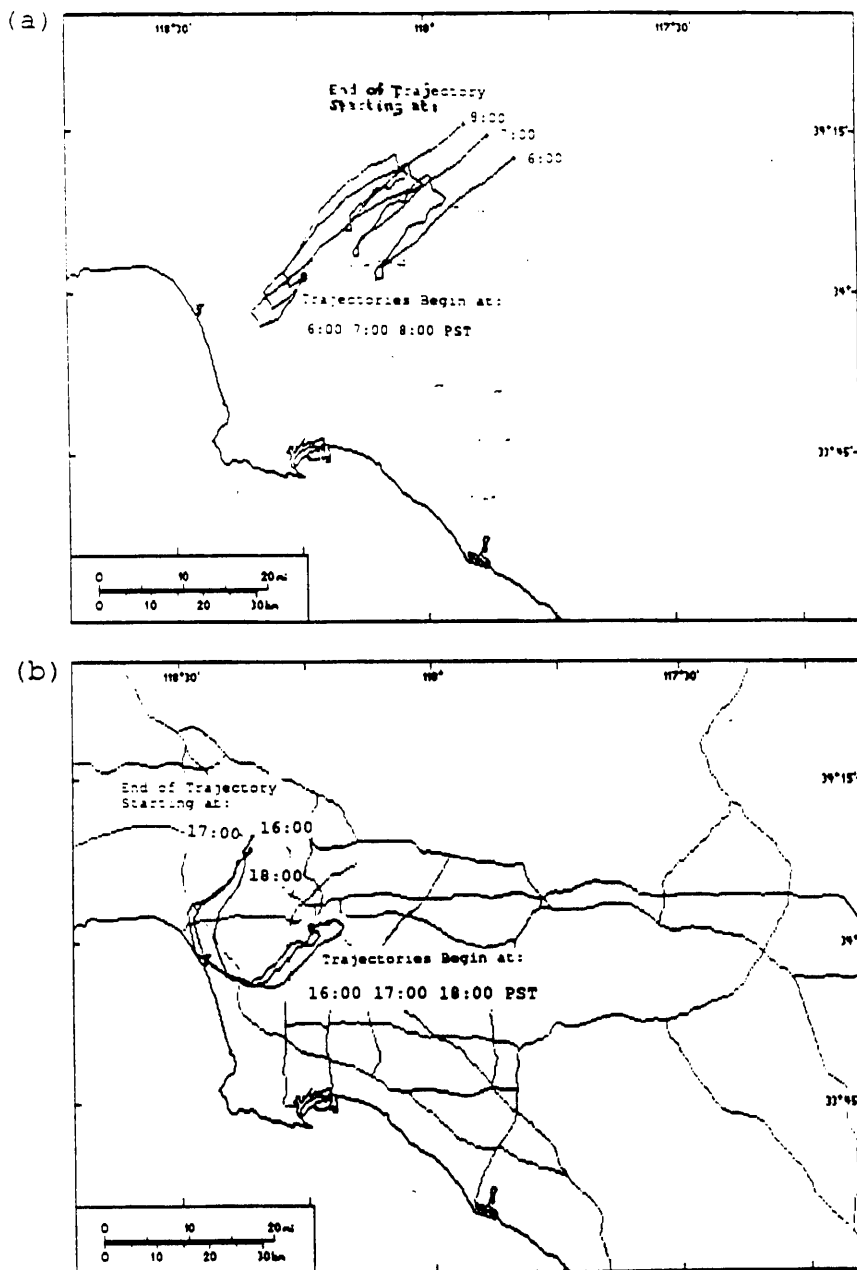


Figure 8. Surface wind trajectories from Vernon starting at (a) 6, 7, and 8 PST lasting 36 hours on November 12-13, 1987, and (b) 16, 17, and 18 PST lasting 24 hours on December 10-11, 1987.

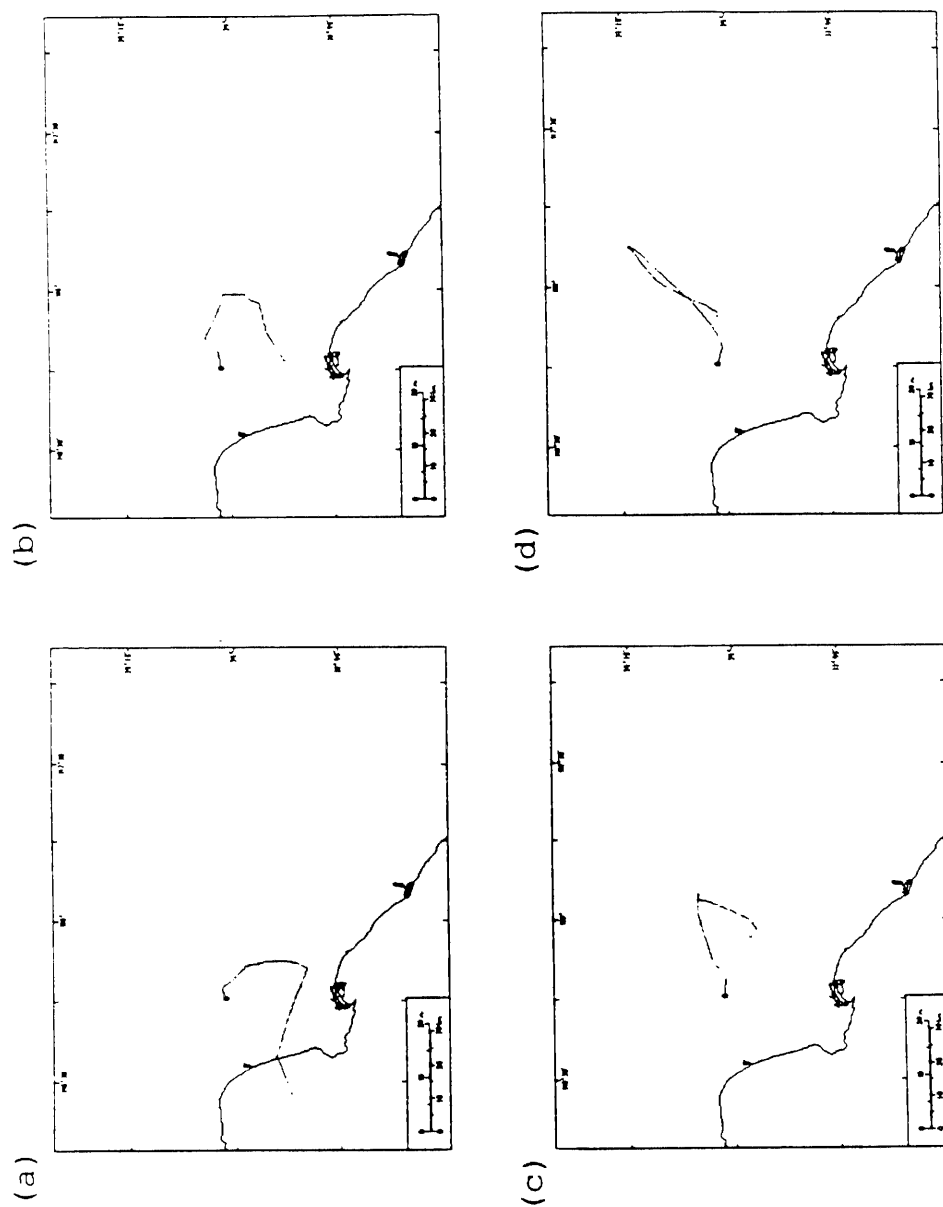


Figure 9. Upper air wind trajectories from Vernon starting on November 12, 1987, starting at 6:00 PST and lasting 18 hours at altitudes AGL of (a) 100 meters, (b) 200 meters, (c) 300 meters, (d) 400 meters.

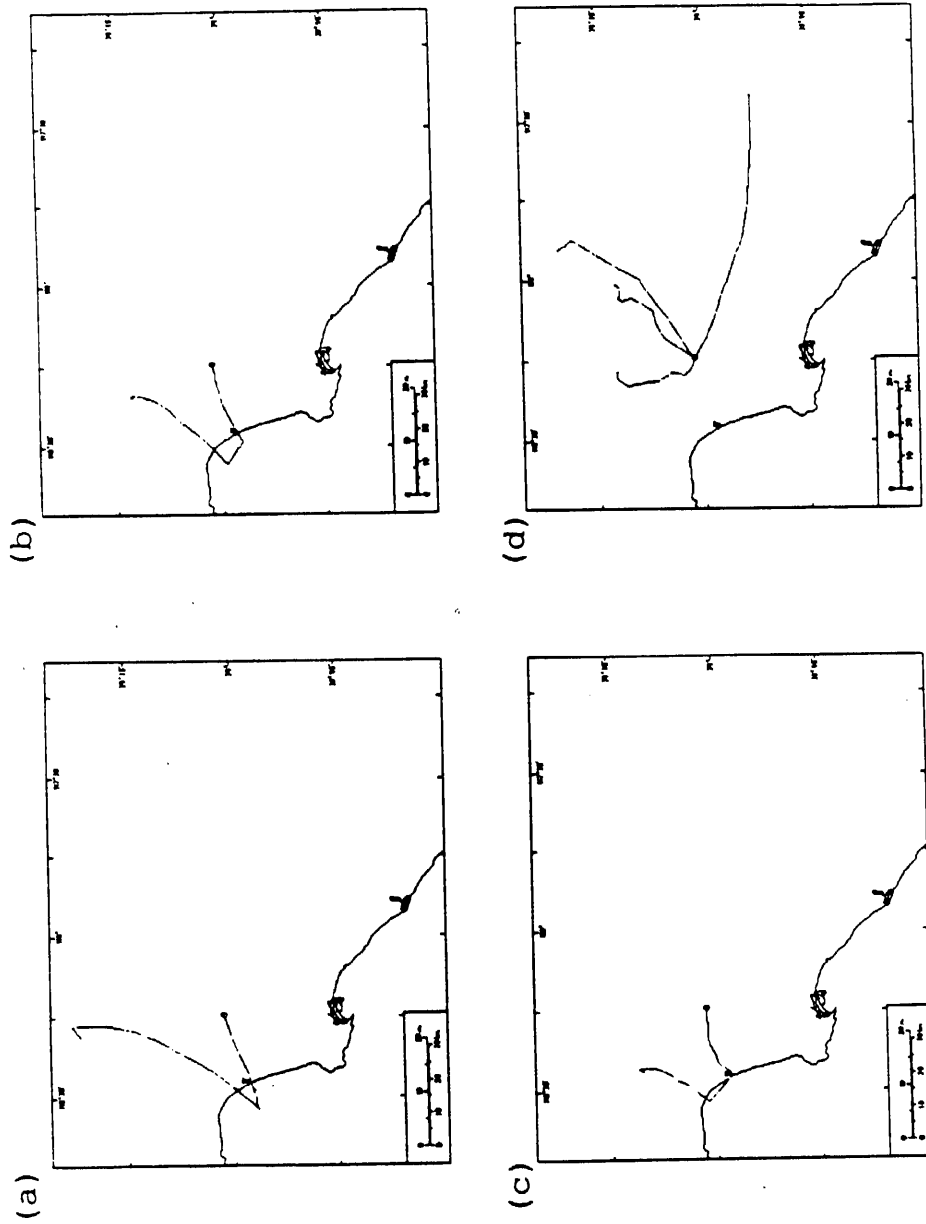


Figure 10. Upper air wind trajectories from Vernon starting on December 10, 1987, starting at 16:00 PST and lasting 18 hours at altitudes AGL of (a) 200 meters, (b) 300 meters, (c) 400 meters and (d) 600, 800, 1000, and 2000 meters going clockwise about Vernon.

2.2 Establishing Upper Bounds on Vertical SF₆ Tracer Transport During the 1987 SCAQS

Abstract

This document briefly describes how the commonly occurring temperature inversions in the Los Angeles Basin were used to establish upper bounds on the vertical tracer transport. Upper bounds on the transport of sulfur hexafluoride, SF_6 , tracer during the Southern California Air Quality Study were obtained by examining the coinciding vertical atmospheric structure. Results are tabulated for the SF_6 experiments performed during July, November, and December, 1987.

Introduction

Within the Los Angeles Basin the vertical atmospheric structure is complex and varies with horizontal position as well as in time. The vertical transport and dispersion of surface-based emissions that results from the dynamic micro-meteorological conditions and winds within the basin has been studied extensively. For instance, Edinger (1959, 1973), Lea (1967), Stephens (1975), and Blumenthal et al. (1978) among others, have reported and used the vertical atmospheric structure of the Los Angeles Basin to explain some observations made during smog episodes.

Ground and near-ground released gaseous tracers, like other surface-based emissions, are influenced by the same transport mechanisms. Consequently, the closure of a tracer mass balance in the vertical direction can directly benefit from a quantitative description of the lower troposphere. This section describes how such descriptions gathered during the 1987 Southern California Air Quality Study enabled mass balancing on the SF_6 tracer data from the same period.

Typically, tracer mass balances are done to establish the fraction of tracer encountered at particular times and locations during a study. Reible (1982) has demonstrated that the fractional recovery can be determined within a 25 percent error given typical uncertainties for the pertinent data. Since tracer measurements are often obtained by ground based sampling, and airborne sampling techniques may be impractical for a variety of reasons, some method must be used to determine the vertical extent of tracer transport. In lieu of elevated tracer measurements, the extent of vertical tracer transport is inferred from plots of the vertical atmospheric measurements. Once the extent of vertical transport is estimated, elevated concentrations are extrapolated from the tracer concentration measurements at the surface. This particular step is generally associated with a large uncertainty, possibly 40 percent or more. Minimization of this er-

ror is desirable and can be accomplished if a detailed hourly description of the lower troposphere's structure is available. To understand how the atmospheric structure is used in the analysis, it is important to be familiar with some practical aspects of doing the mass balance. For this reason some readers may find the next six paragraphs informative.

Two general methods of calculating atmospheric mass balances directly from measurements have been used. The first method is to measure the flux of material passing through a stationary point, line, or preferably a surface, and integrate the flux over time (Reible 1981, Shair 1982).

$$m_t = \rho \int \int C(x, y, z, t) \underline{\mathbf{v}}(x, y, z, t) \cdot \underline{\mathbf{n}} \, dA \, dt \quad (2.2.1)$$

Equivalently, Equation 2.2.1 is applied to actual measurements in its differential form as Equation 2.2.2 below:

$$m_t = \rho \sum_{i=1}^{i=n} C_i(x, y, z, t) \underline{\mathbf{v}}_i(x, y, z, t) \cdot \underline{\mathbf{n}}_i \, \Delta A_i \Delta t. \quad (2.2.2)$$

In Equations 2.2.1 and 2.2.2, m_t is the calculated total mass of tracer in kg , ρ is the tracer density in kg/m^3 , and $C(x, y, z, t)$ in m^3 tracer/ m^3 air is the time dependent tracer concentration in a fixed Cartesian coordinate system relative to the ground. For convenience, the directions x and y are in the horizontal plane of the ground, and z is the vertical direction. The unit vector $\underline{\mathbf{n}}$ is normal to the control surface, A . The vector velocity is denoted by $\underline{\mathbf{v}}$. In the differential form the i denotes an individual measurement. For typical measurements in the Los Angeles Basin, the principal component of the wind is horizontal, suggesting that ideally the control surface, A , should be perpendicular with the ground. Tracer measurements are then made at points along a line on the ground. Optimally, these points should also be perpendicular to the direction of flow.

The second method of conducting a mass balance is to simultaneously measure the concentration at many points in an enclosed volume, such as was done in the paper by Horrell et al. (1989). The total mass of tracer observed at time t is determined by Equation 2.2.3, below:

$$m_t = \rho \int \int \int C(x, y, z, t) dx dy dz. \quad (2.2.3)$$

Equivalently, Equation 2.2.3 is written in its differential form as Equation 2.2.4 below:

$$m_t = \rho \sum_{i=1}^{i=n} C_i(x, y, z, t) \Delta x_i \Delta y_i \Delta z_i. \quad (2.2.4)$$

The total mass is found by measuring the tracer concentration on a two-dimensional ground-level surface and making an estimate of the vertical concentration profile. The differential area, and volume of Equation 2.2.2 and 2.2.4 respectively, are calculated based on the physical spacing between the measurement sites.

The above two methods may be best suited to different circumstances. The first method is well suited to conditions where: (1) all the tracer is expected to cross a particular line some time after the release and (2) restrictions exist on the location of sampling sites. This first method was successfully applied to tracer which was advected offshore over the ocean by night, and then measured at the coastline as it returned onshore by day. The second method is probably best used under diagnostic conditions where (1) the direction of tracer transport is not anticipated *a priori*, and (2) the ground-level measurements can be taken on a two-dimensional surface.

Particularly relevant to determining the vertical extent of surface-based emission transport within the Los Angeles Basin is the airspace from ground level to roughly fifteen hundred meters. The mountains which bound the basin,

the sea breeze and a commonly occurring temperature inversion create conditions which may confine the surface based emissions below this height. The existence of a temperature inversion, the height at which it is first observed (its base), and the height at which the atmosphere returns to an adiabatic lapse rate (its top) all affect the ability of this interceding stable air mass to restrict vertical transport. The distance from the ground to the base of the inversion is referred to as the mixing depth or mixing height. Refer to Figure 1, reproduced from Fundamentals of Stack Gas Dispersion (Beychok 1979), to see an idealized rendition of the vertical atmospheric temperature structure. If an inversion exists it is likely to follow a diurnally rising and falling pattern. To identify the inversion and determine the mixing height, among other things, a number of quantities are measured. These include the state variables, temperature and pressure, the moisture content as dew point and relative humidity, and the wind speed and direction. By plotting these measurements and quantities calculated from them, the top and bottom height of the inversion and the approximate atmospheric stability can be determined by inspection.

Briefly, the stability of the atmosphere between the ground and the base of the temperature inversion determines the effective rate of turbulent transport of surface emissions up to that elevation. Beneath the inversion the air is likely to be neutral to very unstable. Beyond this elevation the strongly suppressed mixing within the inversion acts to dampen turbulence and decrease the effective diffusive transport by orders of magnitude. Under these conditions the air is considered stable. Extensive research has been done to estimate vertical diffusivities from meteorological parameters under a wide range of stabilities. A summary of work on this subject which also includes extensive details on its application to air-shed modeling for the Los Angeles Basin was compiled by McRae, Goodin, and Seinfeld (see EQL, 1982). For the purposes of the

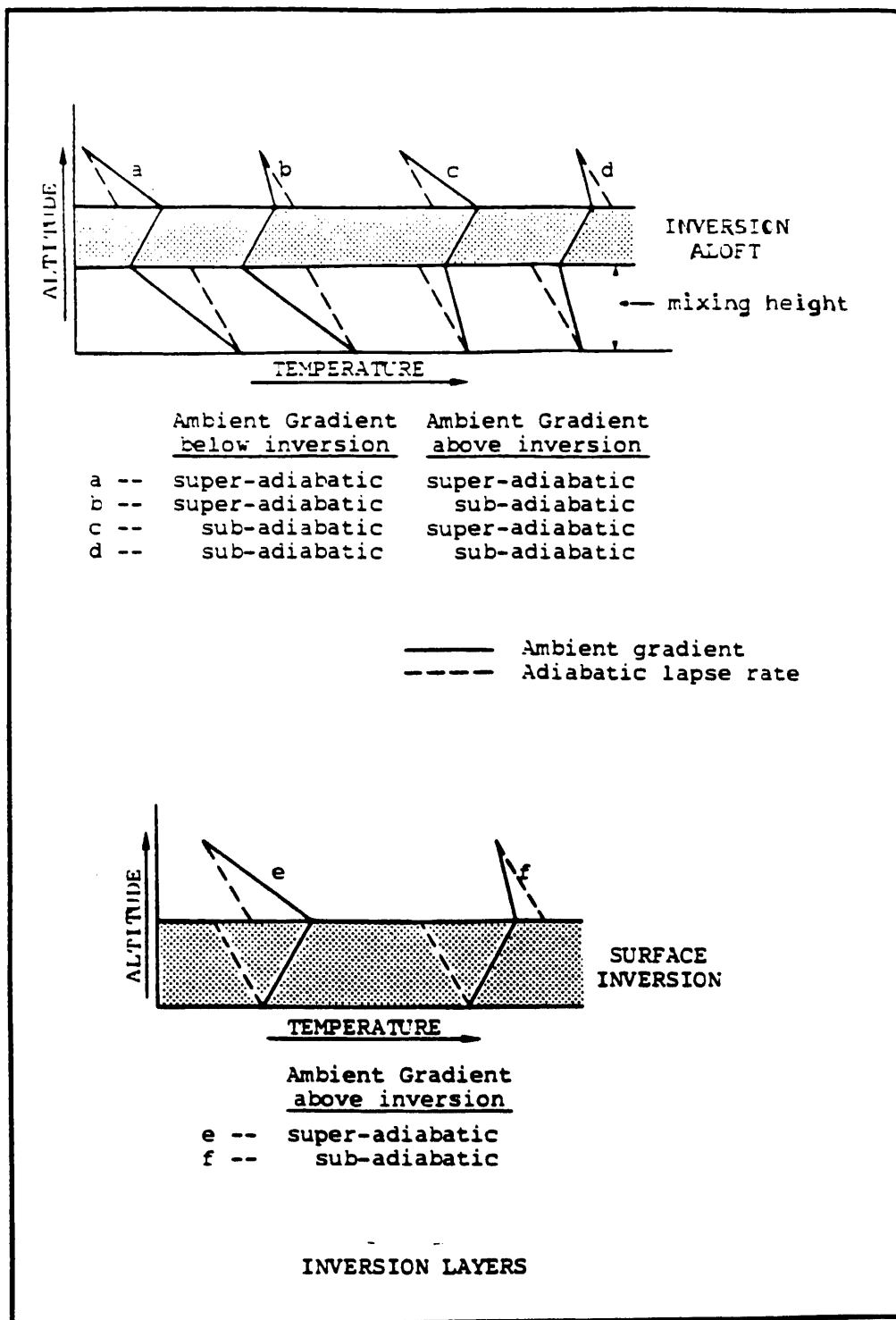


Figure 1. Idealized vertical temperature profiles (reproduced from Fundamentals of Stack Gas Dispersion, 1979).

preceding paper's mass balances, a very simple assumption about the vertical mixing rates was used when an inversion existed. It is that vertical mixing is fast between the ground and the inversion base relative to horizontal tracer transport. This has the equivalent effect of assuming the tracer is constant in the z -direction at the measured ground-level concentration up to the base of the inversion.

For a given time, a reasonable upper bound on the extent of the tracer's vertical transport is the greatest height that the tracer could have been transported to at all previous times since its release. When a temperature inversion is present, the mixing height is the best upper bound. This is an especially good estimate when the inversion height has been constant or monotonically increasing since the tracer release. Under these conditions the air beneath the inversion is likely to be neutral to unstable, and tracer is being mixed readily upward from ground level. When the inversion height has been fluctuating or decreasing, the maximum height since the release is a more appropriate choice. First, the tracer may have been transported to a maximum height at an earlier time when the inversion base was higher. Finally, as the inversion base falls, it may be incorporating tracer-rich air from beneath it. By this mechanism tracer would become trapped in the inversion, and estimates of its maximum vertical dispersion would be too low, based on the current inversion height. This trapping phenomena has been observed with pollutants such as ozone (Edinger 1973). Alternatively, if the inversion base rises or the inversion is entirely destroyed by the eroding effect of unstable air from below (or other mechanisms), an upper bound on the transport may be harder to determine. In such cases the atmospheric stability can be used to compute an upper bound. Similarly, when the inversion begins at the surface and stays there throughout a tracer experiment, bounding the extent of vertical mixing is also done using the atmospheric

stability.

During the Southern California Air Quality Study, three of the four tracer releases began in the morning. Of these, two were done in the summer and on days where an elevated inversion existed. The third morning release was in the fall period, and the fourth release was a late afternoon fall release. Both were associated with neutral to stable conditions where the inversion bases were much lower than the summer releases.

The vertical measurements from the Southern California Air Quality Study will be presented below. The purpose of this section is to present the data which is available from the SCAQS, and at the time of this research had not been reported nor analyzed in the literature. It is not the purpose of this section to analyze this data in general, but simply to tabulate estimates of mixing heights, and clearly document how such estimates were obtained. These tabulated mixing heights and the preceding discussion are the basis for closing the mass balances described in the paper entitled "Ground Release SF_6 Tracer Experiments Used to Characterize Transport and Dispersion of Atmospheric Pollutants During the Southern California Air Quality Study of 1987" (Horrell et al. 1989). The extensive figure set in Appendix A2.2 is included because it is the best, most comprehensive simultaneous display of these data known to exist.

Method

The data set containing the vertical atmospheric measurements was obtained from the California Air Resources Board for the SCAQS days when sulfur hexafluoride tracer experiments were done. Recall, these days are July 15, September 3, November 12, and December 10-11, all of 1987. The measured quantities in the data set are temperature and dew point in degrees *C*, pressure in *mbars*, relative humidity, wind speed in *m/s*, and wind direction in *degrees off north*. Each data file consists of the measurements taken at one of six times during the day, generally around the hours 5, 8, 11, 14, 17, and 22 PST. Data was reported at 30 to 100 meter intervals sometimes to heights of 5000 meters above ground level (AGL).

Some data sets were lacking in entire categories, i.e., wind speed, or contained a large number of missing data. In such circumstances the mixing heights were reported but the plot was omitted from Appendix A2.2. Furthermore, to reduce the number of plots, only those locations that were impacted by tracer or surrounded impacted locations are presented. The diskette containing the data for November 13, a tracer sampling day, had failed at the time of plotting, but was used at an earlier time to compile the data.

The measured data was used to calculate the potential temperature, the gradient form of the Richardson number (Equation 2.2.6), and the change in wind direction with height at each vertical height that data was reported. The processing program, UPPR:UPPer air vieweR (developed at California Institute of Technology by the author), then simultaneously generated plots for both measured and calculated variables. The potential temperature was calculated using the relationship:

$$\Theta = \Gamma T + T_0. \quad (2.2.5)$$

In Equation 2.2.5, Θ is the potential temperature, Γ is the atmospheric lapse rate, T is the measured temperature, and T_0 is the ground temperature. A constant value for Γ of 0.01 degrees C/m was used. This neglects the effect of water vapor on the lapse rate, but the effect is small (Seinfeld, 1975) and at no time did neglecting it prevent the inversion base from being located. The stability of the air below the inversion base was estimated from the slope of the vertical plot using Figure 2 as a guideline. In addition, a form of the gradient Richardson number and the change in wind speed with height was calculated in order to qualitatively, point-wise, compare their magnitude with the potential temperature and measured quantities. In theory the inversion base should coincide with a large gradient Richardson number because of the sharp temperature (density) change associated with the inversion. It was calculated based on the differential form of its definition.

$$Ri_g = \frac{\frac{g}{T} \frac{\partial \Theta}{\partial z}}{\left(\frac{\partial \bar{U}}{\partial z}\right)^2} \equiv \frac{\frac{g}{T} \frac{\Delta \Theta}{\Delta z}}{\left(\frac{\Delta \bar{U}}{\Delta z}\right)^2} = \frac{g}{T} \Delta \Theta \Delta z (\Delta \bar{U})^{-2} \quad (2.2.6)$$

In Equation 2.2.6, g is the gravitational constant, and \bar{U} is the time averaged speed as a function of z . The other variables have the same meaning as in Equation 2.2.5.

Limitations were imposed on the data and calculated values to ensure that the plots would be free of invalid data and easy to read. The Richardson number was limited to values below ten thousand. Negative values were permitted but seldom observed. Invalid speeds cause calculation of the Richardson number to be suppressed. The change in direction with height was limited to values of ± 18 degrees. Values which exceed these criteria were truncated.

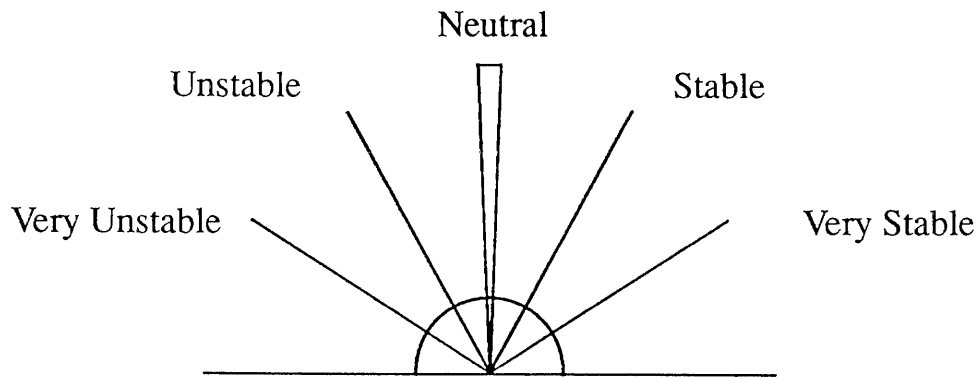


Figure 2. Slope of the potential temperature with increasing height versus range of atmospheric stability.

Results

The estimated mixing heights are presented in Tables 1-3. These values were obtained directly by interpreting Figures 1-68 of Appendix A2.2, and some additional plots not reproduced here, showing vertical data profiles upto 1500 meters. The estimated heights correspond with reasonable recovery rates for the mass balances.

Acknowledgements

I would like to thank Don Lehrman of Technical and Business Systems for providing me with some preliminary data and the data format just following the summertime SCAQS. This allowed me to develop the analysis tool UPPR for creating and manipulating the data well in advance of the release of the verified data by the California Air Resources Board in 1988. I would also like to thank Chuck Bennett of the California Air Resources Board for providing me with the verified data.

Date	Site Name	Hour of the Day	Inversion Base Height (meters)
July 15	Burbank	5	700
		8	450
		11	550
		15	500
		17	550
		22	500
	El Monte	5	550
		8	550
		11	375
		14	375
September 3	Burbank	18	550
		5	175
		8	175
		11	300
		14	300
		17	375
	El Monte	22	600
		8	175
		12	300
		14	750
		17	500

Table 1. Mixing Depths During the Summertime SCAQS 1987

Date	Site Name	Hour of the Day	Inversion Base Height (meters)
November 12	Burbank	4	0
		8	150
		11	150
		14	375
		17	0
		22	0
	El Monte	6	0
		11	375
		14	750
		17	0
	Long Beach	5	0
		8	0
		11	375
		14	200
		17	200
		22	0
	LMU	5	0
November 13	Burbank	5	275
		8	500
		11	450
		14	750
		18	750
		22	375
	El Monte	7	500
		8	500
		11	600
		14	750
	Long Beach	8	750
		11	750
		14	800
		17	800
	LMU	5	400
		8	700
		11	750
	Ontario	5	0
		9	0
		11	375

Table 2. Mixing Depths During the November SCAQS 1987

Date	Site Name	Hour of the Day	Inversion Base Height (meters)
December 10	Burbank	11	0
		14	375
		17	0
		22	0
	El Monte	5	0
		8	0
		11	100
		15	300
	Long Beach	11	200
		14	400
		17	200
		22	0
	LMU	0	0
		14	200
		17	0
December 11	Burbank	5	0
		8	0
		10	0
		14	200
	El Monte	8	0
		11	200
		17	0
	Long Beach	5	50
		8	50
		11	50
		14	200
	LMU	5	0
		8	0
		11	200
		14	200

Table 3. Mixing Depths During the December SCAQS 1987