ASCOT

ATMOSPHERIC STUDIES IN COMPLEX TERRAIN

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**A COLLECTION OF PAPERS BASED
ON DRAINAGE WIND STUDIES
IN THE GEYSERS AREA OF
NORTHERN CALIFORNIA: PART I**

Marvin H. Dickerson
(Editor)

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FOREWORD

Participants in the DOE ASCOT program met in Gettysburg, Pennsylvania, on April 15-18, 1980, for discussions related to the analysis, interpretation, and modeling of data collected in The Geysers area of northern California. These data were collected during a two-week period in July 1979 and represent an initial field experiment to study nocturnal drainage winds. Two major objectives of this experiment were: (1) to investigate and become familiar with the logistical problems associated with field studies in the area, and (2) to obtain measurements of wind and temperature fields for use in designing a major experiment in September 1980.

Both objectives were met successfully, and many results of the participants' research using data from the experiment are described in the collection of papers that follows. The papers are arranged in order from data analysis and interpretation to modeling.

We hope that the papers presented in this report will give other investigators new insights into the phenomena of nocturnal drainage winds and encourage them to use the data. These data will be available in report form by approximately January 1, 1981.

Marvin H. Dickerson
Scientific Director
ASCOT Program

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LARGE SCALE METEOROLOGICAL INFLUENCE DURING
THE GEYSERS 1979 FIELD EXPERIMENT

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ABSTRACT

A series of meteorological field measurements conducted during July 1979 near Cobb Mountain in Northern California reveals evidence of several scales of atmospheric circulation consistent with the climatic pattern of the area. The scales of influence are reflected in the structure of wind and temperature in vertically stratified layers at a given observation site. Large scale synoptic gradient flow dominates the wind field above about twice the height of the topographic ridge. Below that there is a mixture of effects with evidence of a diurnal sea breeze influence and a sublayer of katabatic winds.

The July observations demonstrate that weak migratory circulations in the large scale synoptic meteorological pattern have a significant influence on the day-to-day gradient winds and must be accounted for in planning meteorological programs including tracer experiments.

I. SYNOPTIC METEOROLOGY DISCUSSION

Although a preliminary survey of the five experimental nights of the ASCOT field program at Anderson Creek Valley, California indicates a basic

similarity in the drainage wind structure from one night to another, the detailed analysis of wind structure reflects significant differences. One potential cause of the difference is the variation in free stream winds at and above the ridge top due to an evolving synoptic scale pressure pattern. Although the mean pattern of high pressure over the eastern Pacific Ocean and low pressure over the southwestern U.S. persisted throughout the period, a series of weak migratory circulation systems accounted for a continually changing gradient wind over the experimental area. This gradient wind exerted a major influence on the observed winds in the more exposed areas of the ridge-valley complex.

The early part of the experimental period, July 18th, 19th and 20th, was dominated by a particularly well developed thermal low pressure area covering the southwestern U.S. Large scale winds were light and temperatures were abnormally high. During the experiment night of July 20-21 a cold front moved through the area from the north and produced thunderstorms in the northern California coastal mountains. A weak high pressure ridge moved into the area after the front passed. During the experiment night of July 22-23 the ridge weakened leaving a weak cyclonic circulation in northern California at lower tropospheric altitudes. Pressure gradients weakened further on July 24-25. Low level divergence developed in central California during the night with convergence into three weak low pressure centers in eastern Oregon, New Mexico, and in the Pacific about 300 miles offshore from southern California. The Pacific low drifted onshore and was present in the Geysers area during the night of July 26-27. Except for the frontal passage of July 20-21, none of the circulation features described here was vigorous enough to produce clouds or weather events. The gradient winds did vary in a continually evolving pattern due to the migratory features in the pressure field. The altitude level at which gradient flow exerts an optimum influence on local topographic wind patterns is the effective ridge height where the terrain is most exposed to the outer flow. Lee side effects of sheltering, separation, vertical streamline confluence and waves will be initiated by

interactions at ridge level. The topography in the Anderson Creek Valley area reaches 975 to 1430 m MSL. The 850 mb level is representative of these altitudes. For purposes of interpretation of interactions of external wind fields with locally generated drainage winds, a set of 850 mb contour charts and streamlines is provided in Figures 1 through 8.

II. MESOSCALE INFLUENCE

The confluence of low altitude air into a summer time thermal low pressure area in the Great Basin represents a characteristic monsoon type of circulation depicted in Figure 9. This pattern is distinctly seasonal but does not vary diurnally. However, superposed sea breeze effects do produce a diurnal pattern of winds close to ground level. Fosberg and Schroeder (1966) document the penetration of marine air into the Middletown, California area. The marine air is significantly modified during its overland travel through the coastal mountains. At Middletown, high temperatures and low humidity are more characteristic of continental air. Wind hodographs derived from 45 summer days reflect a diurnal wind pattern with southwesterly flow peaking in mid-afternoon and a brief period of northeasterly winds after sunrise that Fosberg and Schroeder suggest is due to a land breeze.

The properties of the sea breeze depend somewhat on synoptic features. Fosberg and Schroeder suggest that when the Pacific high pressure cell extends onshore into the Pacific northwest the seabreeze is slowly moving and shallow with a well defined sea breeze front that does not extend far inland. When a weak trough exists over the coastal area the sea breeze is deeper and moves more rapidly and further inland. However, the boundary is more poorly defined than the warmer, shallow case.

Meteorological measurements made close to ground level often reflect a combination of phenomena that may complicate their interpretation. Especially in complex terrain, measurements to be related to the large

scale pressure gradient should be collected above the localized influence of mountain-valley winds or obstacle flows. During the July 1979 Geysers field program LASL operated a dual-theodolite tracked pilot balloon wind profile system equipped with a portable temperature sonde. These measurements extended to 2 to 3 km above local terrain. Figures 10 through 13 represent time-height cross sections of wind direction from the surface to the upper extent of the soundings. The wind directions above twice the ridge height generally agree with the 850 mb streamlines derived from the NWS limited fine mesh analysis. The 850 mb level is the nearest standard analysis level to the ridge top free stream wind and is located at about 1 km above the base of the soundings.

Below about twice the ridge height the winds appear to be affected by local influences. This lower layer has several subregions defined by characteristics of the wind direction cross section. A shallow and generally persistent layer of west to northwest wind is identified with the locally driven katabatic wind. In the transition layer from 200 m to 1 km above ground the wind direction exhibits one or two excursions during the course of the night from westerly through northerly. The largest meanders result in easterly winds for 1 to 2 hours at a time. Above the transition layer the wind directions are steady and agree with the synoptic scale streamlines.

The temperature field in the lowest 1400 m is presented in time-height cross sections (Figs. 14-17) that show several consistencies and some major differences from night to night. All the figures show the development of a shallow stable layer due to nocturnal cooling at ground level after sunset. A zone of warm air persists between 200 m and 600 m (ridge top) until about 0300 PST. The data are insufficient to establish a rigorous relationship but there appears to be a connection between the maximum cooling aloft and a pre-dawn surge of easterly winds. Above 1 km the temperatures generally decrease with elevation and exhibit small but observable oscillations in time.

III. SUMMARY

At the first level of detail, the summer large scale meteorology of the northern California coastal mountains is governed by a monsoon type of flow between the Pacific high pressure cell and a thermal low in the Great Basin of the western U.S.. However, weak migratory circulation systems produce day to day variations of wind direction above the level of local influence. Sea breeze and local terrain dominated winds are superposed on the varying synoptic pattern and control the winds in the lowest 1-2 km above sea level.

A series of soundings that extend above 2 km near the site of a boundary layer meteorological experiment provides a link between the local domains of wind flow and the free stream wind. During the July 1979 Geysers field experiments the LASL pilot balloon soundings with minisonde temperatures fulfilled this interpretive link.

REFERENCE

Fosberg, M. A. and Schroeder, M. J., "Marine Air Penetration in Central California", J. Appl. Meteorol., 5, 573-589, 1966.

850 mb ANALYSIS STREAMLINES FOR 79071900Z

Fig. #1

850 mb ANALYSIS STREAMLINES FOR 79071912Z

Fig. #2

850 mb ANALYSIS STREAMLINES FOR 79072100z

Fig. #3

850 mb ANALYSIS STREAMLINES FOR 79072112z

Fig. #4

Fig. #5

Fig. #6

Fig. #7

Fig. #8

Fig. 9 Resultant surface streamlines for July (after NAVAIR, 1966).

Fig. 10 - Time-Height Cross Section of Wind Direction for July 18-19, 1979

Fig. 11 - Time-Height Cross Section of Wind Direction for July 22-23, 1979

Fig. 12 - Time-Height Cross Section of Wind Direction for July 24-25, 1979

Fig. 13 - Time-Height Cross Section of Wind Direction for July 26-27, 1979

Fig. 14 - Time-Height Cross Section of Temperature for July 18-19, 1979

Fig. 15 - Time-Height Cross Section of Temperature for July 22-23, 1979

Fig. 16 - Time-Height Cross Section of Temperature for July 24-25, 1979

Fig. 17 - Time-Height Cross Section of Temperature for July 26-27, 1979

EXPERIMENTAL METHODS AND PRELIMINARY ANALYSES OF DRAINAGE FLOW OBSERVATIONS DURING THE FIRST ASCOT FIELD STUDY*

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1. INTRODUCTION

The Atmospheric Turbulence and Diffusion Laboratory (ATDL) conducted several types of experiments during the first ASCOT** field study at "The Geysers" geothermal energy production site in northern California, July 15-27, 1979. The object of these preliminary studies was to gain some insight into the characteristics of nocturnal cold air drainage in this complex terrain. Such understanding is important for air quality studies, since present (and many future) geothermal steam releases are generally close to the surface, within the zone of influence of drainage flows. Contaminants emitted with the steam, such as hydrogen sulfide and heavy metals, could conceivably be carried downhill to populated regions in this resort area.

Nappo et al. (1980) discuss the trajectories of neutrally buoyant balloons released into the cold air drainage along the Anderson Creek watershed. The

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** Atmospheric Studies in Complex Terrain.

present paper describes wind and temperature data from two instrumented masts and two wiresondes operated in the same general area, and concludes with some hypotheses about the observed flow patterns. Experiments were conducted from before sunset to after sunrise on several occasions; only results from the nights of July 22-23 and 24-25 are presented here. A comprehensive summary of the experimental techniques and data from all the ATDL sites may be found in Nappo et al. (1979).

2. EXPERIMENTAL SITES AND INSTRUMENTATION

Two telescoping masts (8.7 m height) were deployed along the Anderson Creek watershed above the hamlet of Anderson Springs. The upper site was near the confluence of Gunning and Anderson Creeks at an elevation of about 628 m MSL. The lower site was roughly 400 m north of the eastern end of the settlement at an elevation of about 466 m MSL; this site was fairly close to the junction of Anderson and Putah Creeks (Figure 1).

The upper mast was erected in a grassy clearing on a narrow, nearly level promontory extending some 200 m to the SSE between Anderson and Gunning Creeks. The clearing was at the foot of a steep slope rising perhaps 25 m to a 40 m wide flat along the SE side of Socrates Mine Road. The bed of Anderson Creek was about 70 m below the end of the promontory. The mast was fitted with four low-threshold cup anemometers (Thorntwaite model 106) at heights of 1.2, 3.0, 5.1, and 8.6 m above ground. The anemometers were individually calibrated beforehand in the ATDL wind tunnel. Data from the anemometers were recorded on electromechanical counters. A standard darkroom timer was used to interrupt the counting automatically and audibly signal the end of a measurement period. An observer then recorded the counts from each anemometer in that period, reset the

timer, and restarted the count. An interval of one hour was used for most runs, although shorter and longer periods were utilized on some occasions. Calibration curves were used to deduce the average wind speed at each level for the measurement interval. The resulting wind speeds are believed to be accurate to about ± 0.1 m/s.

A low-threshold wind vane (Thorntwaite WD-1) was mounted at 8.6 m. Its potentiometer was energized by a regulated power supply, and the wind direction was sampled every ten seconds by a miniature automatic data logger (Campbell Scientific CR21). Three-minute averages of these readings were then written on to a magnetic tape cassette. Accuracy was limited mainly by the orientation of the vane to the north in the field; this alignment is believed accurate to within $\pm 15^\circ$.

Six linearized thermistors (YSI model 44202) were mounted in naturally aspirated solar radiation shields patterned after a Canadian design (McTaggart-Cowan and McKay, 1976), and placed about 0.5 m out from the mast at elevations of 1.2, 2.2, 3.1, 5.1, 7.1, and 8.4 m above ground. The thermistors were operated in a Wheatstone bridge arrangement powered by an extremely stable supply (2.500 V, to avoid self-heating). These data were also read every ten seconds by the automatic data logger, and three-minute averages were written on the cassette tape. Before and after the experiment, the thermistors were individually calibrated; the raw data are believed to be accurate to within $\pm 0.2^\circ\text{C}$, excluding the radiation shield-induced errors. Application of individual calibration curves permitted a laboratory accuracy of $\pm 0.04^\circ\text{C}$; in view of the many uncontrolled variables in the field work, however, the corrected data are probably accurate only to $\pm 0.1^\circ\text{C}$, again exclusive of shield errors. McTaggart-Cowan and McKay's (1976) work suggests that errors in nocturnal observations with these radiation shields should be quite small (-0.04°C for a 1.5 m/s wind), so no corrections for this effect have been applied.

The lower mast was located in an open grassy field some 200 m south of Socrates Mine Road and about 750 m west of the intersection of State Highway 175 and Anderson Springs Road. The field sloped off to the south toward the wooded area along Anderson Creek; the view to the northwest was roughly up this watershed, although the upper mast site was not visible. To the east, the field dropped off toward Putah Creek, the principal stream draining this region. This mast was fitted with a single Thornthwaite anemometer and a Thornthwaite wind vane, both located at the 8.6 m level. Data recording and reduction techniques were the same as for the upper mast, and the same accuracies should apply. The mast was also equipped with six radiation-shielded thermistors, identical to those at the upper site. The sensors were 1.2, 2.2, 3.1, 5.1, 7.1, and 8.4 m above ground. Recording was again by automatic data logger; frequencies and accuracies were as stated above.

In addition to the instrumented masts, two tethered balloons were used to collect detailed temperature structure information above each site. A 4.25 m³ "blimp" (A.I.R. model TS-1BR-2) carried a linearized thermistor aloft. The circuitry and calibrations were identical to those used on the masts, except that only a simple "stacked plate" solar radiation shield was used. The maximum altitude of the balloons was limited to about 61 m by the length of the wires connecting the thermistor to the ground-based electronics package. Temperature was read directly from a digital panel meter and recorded manually. Calibrations before and after the experiment indicated that the readings were accurate to within the resolution of the display, $\pm 0.1^{\circ}\text{C}$, exclusive of the radiation shield errors. The sonde package had a time constant of roughly one minute, which limited the rate at which ascents or descents could be performed.

A hand-powered winch was used to crank the balloon and sonde to the required altitude. Height location was determined from marks at 3.05 m intervals on the

tether line; the balloon was assumed to fly directly overhead, so that slant distance was equal to altitude. Because of the aerodynamic lift generated by this balloon design and the light winds encountered aloft, this appeared to be a reasonable assumption. For angular displacements of less than roughly 25° from the vertical, the altitude determination was accurate to within 10% or better. Observed slant angles were nearly always smaller than 25° .

3. OBSERVATIONS AT THE UPPER SITE

Short walks about the upper site revealed several interesting phenomena. Whenever cold air drainage occurred, the wind was much brisker at the mast site (1.5 to 2 m/s) than on the flat up the hill, where only light breezes were felt. The air felt cool at both of these locations, but a short distance above and west of the flat the air was perceptibly warmer and stagnant. On the far end of the promontory, which was a few meters higher than the mast site, the air also felt warm and there was no noticeable wind. During the early evening, before the drainage flow was fully established, "puffs" of appreciably warmer air were noted at the mast location by several observers. These seemed to be associated with cross-ridge winds from the west, while during periods of steady cool wind, the wind vane pointed directly upslope (roughly NNW) toward the flat above.

Figure 2 shows mean wind speeds at the upper mast during two nights of the experiment. Despite the data gaps it appears that cold air drainage was fully established before midnight in both cases (about 2300 PST on July 22 and 2215 on July 24). Before the drainage began, wind speed increased with height, as is usual for daytime observations. After drainage started, however, the highest wind speeds occurred at the 1.2 and 3.0 m levels, and the lowest speeds were at the top of the mast. The steadiest wind was observed at the lowest level, 1.2 m. These features

suggest that the drainage at this site was quite shallow, occasionally less than the height of the mast. Notice that early in the evening (2000 to 2030) of July 24, the wind profile had already changed from its normal daytime character to a drainage-like pattern. Sonde temperatures observed at this time exhibited a rather marked decrease as the surface was approached. However, the 2030 to 2100 average speeds returned to the daytime format which persisted until 2215. The temperature change over the 3 m to 61 m range was fairly small during this "throwback" interlude. Subjectively cool periods punctuated by warm "puffs" were noted until midnight, when the drainage pattern had finally settled down. One explanation for such behavior is that cold air flow actually began quite early, but was occasionally "pushed" off to the east of the mast by the still-strong prevailing westerly wind. The 2000 to 2030 period perhaps was a lull in this wind, since the average speed at 8.6 m dropped to 1.6 m/s from its earlier values in excess of 2 m/s.

Figure 3 shows some high resolution time sequences of sonde temperatures at selected heights; in two cases, concurrent one-minute-average wind speeds were observed at the 3 m level. Temperatures read every 15 seconds are plotted in Figure 3a to show the temperature fluctuations typically observed; rapid changes of up to $\pm 0.5^{\circ}\text{C}$ were found at the 3 m level, superposed on longer term variations of a degree or so, which were observed at all levels. The records are too short to determine any possible periodicity in these latter excursions, which several drainage flow theories predict (e.g., Defant, 1949; Fleagle, 1950). The long term variations might also be due to "sloshing" of a stable air mass within the surrounding valley "bowl"; since the temperature gradient above the drainage layer was generally somewhere between adiabatic and isothermal, a 0.7°C jump in temperature could correspond to a 70 m temporary subsidence.

Strong temperature/wind correlations are evident in Figure 3b. The temperature at 12.2 m was positively correlated with the 3 m wind; the temperature

dropped by 0.8°C as the wind decreased by about 0.5 m/s and both then began to recover together. Exactly the opposite behavior was seen when the temperature sonde was at 3 m; the windspeed maximum then corresponded to the temperature minimum, and the wind diminished as the temperature rose.

Figure 4 shows a number of vertical temperature profiles obtained from the sonde system on the night of July 24-25. Some caution is needed in interpreting these profiles, in that the results are neither "snapshots" nor time averages — rather like pilot balloon soundings for wind. The profiles required about 12 to 16 minutes to complete, allowing generally one minute at each level for temperature stabilization of the sensor. Within the lowest 10 m, two minute delays were usually allowed in view of the much larger temperature changes between levels at these heights. Substantial changes in the temperature field were therefore probably encountered during any single run, judging by the variations shown in Figure 3.

Profiles from the early evening are shown in Figure 4a. The 1936-1950 descent showed a slightly stable layer near the surface. The 2007-2021 sounding, however, showed a very stable layer below about 12 m, corresponding to the evening's first episode of drainage wind velocity profile (Figure 2b). The apparent breakup of this early drainage profile (Figure 2b) was reflected in the 2042-2055 and 2111-2124 temperature soundings, which were much less stable near the ground. By the 2306-2319 run, however, the drainage was apparently reestablished, although confirming wind profiles were unfortunately not available for this period.

Hourly descents after midnight are shown in Figure 4b. All were extremely stable in the lowest 6 to 9 m with gradients of $70^{\circ}\text{C}/100\text{ m}$ or more. Note the similarity of these profiles to the earlier 2007-2021 and 2306-2319 soundings. The median of these post-midnight profiles was nearly isothermal above 42 m, and increased at about $1.5^{\circ}\text{C}/100\text{ m}$ between 9 and 42 m. Below 9 m, the profiles were all markedly stable, suggesting that the drainage layer thickness was only

about 9 m at this location. The fairly steady cooling seen at upper levels before midnight was replaced in these later runs by seemingly random warming or cooling of roughly 0.5°C , perhaps related to "sloshing," as discussed above. The 3 m temperatures also varied little from midnight to sunrise except for the 0404-0417 run, which was a full degree warmer than the earlier runs. On the other hand, the post-sunrise soundings, 0503-0516 and 0532-0546, were significantly cooler at 3 m than any earlier result.

4. OBSERVATIONS AT THE LOWER SITE

Figure 5 shows the 8.6 m level wind speed at the lower site for the night of July 24-25. Ten-minute (and a few twenty-minute) averages were determined at this site, so the record displays some rapid changes not visible in the hourly-averaged record from the upper mast (Figure 2b). Quite large speed variations occurred in the early evening, superposed on an average speed of perhaps 1 m/s; just about midnight the wind fell off to a mean of about 0.55 m/s which lasted until sunrise, when the speed began to rise again. The short-term fluctuations appear to be periodic, but the gaps in the record prevent verification of this.

Vertical temperature profiles from the lower site are shown in Figure 6 for this same night. These profiles are continuous composites of mast sensor data (below 9 m) and balloon ascents or descents (10 to 61 m) taken over periods of 30 minutes or less, so they must be viewed with some caution. Calibrations done both before and after the field work indicated no problems with the temperature sensors. Nevertheless, large changes in temperature during the early morning hours seemed to coincide with the top of the mast, where measurement shifted from one system to another. Perhaps the different radiation shield designs on the sonde and mast sensors affected the results. Earlier in the evening, however, the changes at the

"crossover" height were not very great. In view of this and the good calibration results cited above, the composite profiles were tentatively accepted. Further testing and side-by-side comparisons of the two instrument systems are planned to clarify this point.

Around sunset, the 1830-1900 ascent showed a slightly unstable near-surface layer with an inversion developing around 10 m, and an adiabatic layer up to roughly 30 m. Above 40 m, the profile was essentially isothermal, a feature which persisted through the night at both mast sites. As darkness fell, the surface layer became increasingly stable, with rapid cooling near the surface, and a smaller cooling rate aloft. After midnight a strong inversion appeared to develop near the top of the mast; during this time, the wind at 8.6 m was from the NNE at about 0.55 m/s (see Figure 5), though winds at the surface were lighter. Assuming the data are reliable, as discussed above, a possible explanation is that cold air drainage from the main watersheds — principally Putah and the combination of Anderson and Bear Canyon Creeks — began to fill up the lower valley basin after midnight, so that cold air began to "back up" into the mast site. The resulting cold air "pool" was nearly stagnant and rather shallow. The very cold layer at the surface may represent a local drainage underlying the "pool" at this location. It is interesting that at this site, too, the coldest temperatures were observed after sunrise as the wind speed began to rise.

Figure 7 indicates the nocturnal cooling rates at 1.2, 8.4, and 12 m. Two distinct cooling regimes appear, with the transition occurring about 0130 PST. The cooling rate was strong early in the night, ranging from 1.6°C/hr at 1.2 m to 1.1°C/hr at 12 m; the corresponding values after the transition were 0.37°C/hr and 0.20°C/hr , respectively. These results may be useful for numerical simulations.

Net radiometer readings at a height of 1 m near this site showed R_N varying slightly from about -58 W/m^2 at 2015 PST to -52 W/m^2 at 2320, to -49 W/m^2 at 0205. These values appear to be consistent with the observed cooling rates and an average drainage layer depth of about 50 m, although a more detailed examination of the surface energy budget would be desirable.

5. INTERPRETATIONS AND CONCLUSIONS

Consideration of the above results, observations of dawn smoke releases high on Anderson Creek, and careful study of topographical maps of the area suggest the following drainage patterns. Since the smoke emitted into the Anderson Creek drainage was mixed to a depth of only 40 or 50 m above the creek bed, at least 10 or 20 m below the upper mast site, it seems clear that the site did not experience this drainage pattern. The topographical map (Figure 1), however, shows a secondary valley along Gunning Creek, which joins Anderson Creek just beyond the promontory where the mast was placed. This valley is only 20 or 30 m below the mast site at its closest location. Furthermore, as Gunning Creek runs downhill above Socrates Mine Road, its valley broadens as the mast site is approached. If the drainage flow depth within this lesser valley is on the order of 20 to 40 m, then the flow can wash over the upper mast site to a depth of perhaps 10 m. The warm "puffs" experienced early in the evening could be produced by a momentary strengthening of the prevailing wind, pushing this Gunning Creek drainage off slightly to the east of the mast. The absence of drainage winds toward the end of the promontory can be explained by the steep dropoff of the ridge, on one side to Anderson Creek and on the other to Gunning Creek. The cold drainage flow moving out along the promontory from the mast site probably just split and fell down these steep slopes to the valleys below before it could reach the end of the ridge.

Very early in the evening (about 2000 PST), the 61 m sounding at the lower mast was about 1.7°C warmer than its counterpart at the upper site; this is quite close to the adiabatic lapse rate, since the sites are about 162 m different in elevation. However, the lower site cooled much more rapidly and continuously than the upper, while the 12.2 to 61 m gradient became much more stable (about $6^{\circ}\text{C}/100\text{ m}$ vs. nearly isothermal) at the lower mast. These effects seem consistent with the hypothesis mentioned earlier that the lower mast was within the "pooling up" region of cold air drainages from Anderson, Bear Canyon, Gunning, and Putah Creeks. These watersheds all join in a rather flat zone just east of the mast, and must exit the area through a fairly narrow pass along State Highway 175. If the outlet is small enough to restrict this pool of cold air, the resulting "puddle" could gradually back up into the lower mast site, producing a cold and stable layer. Cold drainage from creek beds close to the mast might then undercut this layer to produce a highly stable layer in the lowest 10 m or so. The coincidence of significantly colder temperatures and higher wind speeds at both sites after sunrise may be due to a shift of the main drainage patterns, induced by early-morning warming of the slopes across the valley.

Figure 8 illustrates the drainage patterns suggested above. Obviously the present data are insufficient to support such an extensive hypothesis. This speculation seems worthwhile, however, in that it can provide some guidance for additional experimental work during the ASCOT program. In particular, intensive studies of the time-varying depth of the cold air "pool" northeast of Anderson Springs, and a rough mass balance of the incoming and outgoing drainage flows along the major stream beds are recommended. Drainage of air pollutants into a very stable layer of the sort postulated could have potentially significant consequences.

6. ACKNOWLEDGEMENT

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FIGURE 1. Topography of the experimental site. ATDL mast and wiresonde locations are indicated by crosses. Access was via Socrates Mine Road. Major stream are indicated.

FIGURE 2. Hourly-averaged wind speeds from upper mast for two nights, July 22-23 and July 24-25, 1979. Gaps indicate missing data.

FIGURE 3. (a) Temperature sequences at 3.0, 9.1, and 61.0 m; readings every fifteen seconds. (b) Temperature and wind speed, one-minute averages. All data from upper site.

FIGURE 4. Vertical temperature profiles at upper mast for night of July 24-25, 1979. (a) Before midnight soundings. (b) After midnight soundings. Times (PST) indicate beginning and end of a run. Dry adiabat indicated by broken line.

FIGURE 5. Mean wind speed at 8.6 m level, lower mast site, for night of July 24-25, 1979. Broken line indicates missing record.

FIGURE 6. Vertical temperature profiles at lower mast site, night of July 24-25, 1979. Times (PST) indicate beginning and end of a run.

FIGURE 7. Nocturnal cooling rates at lower mast site at height of 1.2, 8.4, and 12.2 m. Lines are best-fits to the data.

FIGURE 8. Tentative nocturnal cold air drainage patterns near the confluence of Anderson and Putah Creeks. The exact depths of the individual flows and of the cold "puddle" east of Anderson Springs are quite uncertain.

A COMPARISON OF ATMOSPHERIC TEMPERATURE STRUCTURE AT THE WALL AND IN THE MIDDLE OF A VALLEY

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1. INTRODUCTION

Several authors (Emslie, 1979; Wilkening and Clements, 1974) have recently found that the measurements of temperatures near ground level on hillsides and valley walls serves as a reasonable approximation to the ambient nocturnal lapse rates. If this result is generally valid in a variety of terrain configurations and meteorological conditions it can offer an inexpensive method for documenting the height, depth and temperature stratification of inversion layers for pollution transport estimates. There are a number of complications that can be perceived to affect the validity of ground level temperatures on hillsides as an estimation of the free air temperature. Local radiative cooling is one while the downslope drainage of cold air is a different but related phenomenon. Both processes are maximized in shallow layers of air close to the ground and their influence may be reduced by the presence of turbulent mixing from an externally imposed wind or by locating the instrumentation on locally exposed promontaries.

At Anderson Creek Valley in the Geysers area of California we had an opportunity to compare temperatures on the valley wall with temperature profiles in the middle of the valley. Temperature profiles were observed on a regular basis on four nights in July 1979 at three locations. Los Alamos Scientific Laboratory released balloon borne sondes at one location while NOAA and Pacific Northwest Laboratory used tether sonde systems at two other sites. The valley was accessible

by road from the base to the ridge. We ran a mobile survey of temperature several times each night using a fast-response instrument mounted on the roof of an automobile.

2. RESULTS

Figures 1-3 are composites of the available soundings and the surface survey temperatures for three runs on each of three experimental nights. The surface temperatures are clearly colder than the air in mid-valley at the same elevation. The scatter in the mobile survey data points reflects the local variations in surface temperature primarily associated with shallow currents of cold air drainage. Patterns of temperature were reproduced in several runs during a given night as well as from one night to another. The coldest temperatures were not necessarily related to the best defined drainage channels, e.g., creek beds. There were several locations of systematically cool temperatures that were not obvious from a simple inspection of a topographic chart. This observation suggests that the cold air drainage paths are determined by some features of the inertia of the moving air and involve some vertical characteristics of the drainage layer. The local temperature variations were smallest early in the evening and the contrasts grew during the course of the night. The early evening runs on July 22 and 24 showed a trend that agreed with the ambient lapse rate while the pre-dawn temperatures reflected a scatter that overshadowed any small trend that might have been present. The night of July 26 was somewhat different. The evening trend in surface temperatures with elevation was less stable than the soundings. The 0300 PST run showed the most stable lapse rates of the entire two week period and the surface-based observations generally agreed with the soundings.

The two additional features of Figures 1-3 reflect the shallowness of the cold air drainage. The triangles are temperatures measured at 10 m above ground on towers operated by Lawrence Livermore Laboratory and Pacific Gas and Electric Company. They tend to be between the surface temperatures and the soundings, but especially after midnight, they are quite close to the soundings. This suggests that the strongest temperature contrasts tend to be confined to a layer shallower than 10 m. The tether sondes that were operated from sites at two different elevations as shown in the figures reflect the strong temperature gradients in the lowest 10-20 meters above local ground level. The soundings agree remarkably well above the height of local ground influence.

The collection of three vertical soundings and several surface-based temperatures along an approximate axis up the slope offers the opportunity to investigate the driving forces for slope winds. Figures 4 and 5 are cross sections of potential temperature (reduced to the altitude of the TV repeater site) along the axis connecting the observation sites. The closely spaced isentropes are confined to a layer less than 100 m in depth, suggesting that the primary forcing of a buoyancy driven drainage wind would be that shallow. Measured wind profiles at the TVR site show evidence of a local velocity maximum at about 60 m. This low level feature did not dominate the profile but was oriented along the valley axis. It is reasonable to view the low level maximum as a buoyancy driven drainage flow phenomena. Typical speeds at the level of the local wind speed maximum were 4 m s^{-1} . It is interesting to estimate speeds that would be generated by the density gradients on the slope to see if they compare in magnitude with the observed drainage component.

3. EVALUATION OF THE CIRCULATION THEOREM

Figures 4 and 5 provide enough information to estimate the strength of the drainage flow by evaluating Bjerknes' circulation theorem:

$$\oint_Q \left(\frac{dv}{dt} \right) \delta s = - \oint_Q \alpha \delta \rho \quad . \quad (1)$$

Equation (1) is discussed by Haltiner and Martin (1957) among other authors. Basically, the theorem says that the line integral of acceleration on an arbitrary closed curve, Q is equivalent to the product of pressure and inverse density integrated along the same curve. On a two-dimensional plane curve Q . Equation (1) can be re-cast into:

$$\oint_Q \left(\frac{dv}{dt} \right) \delta s = R \int_A \frac{1}{\rho} \nabla T \times (-\nabla \rho) \cdot \delta \vec{A} \quad (2)$$

where the right hand side is the area integral of the cross product of the temperature and pressure gradients. If the isobars are parallel to the isotherms, the integral is zero and there is no circulation generated. The isobars associated with the temperature field in Figs. 4 and 5 are horizontal so there is clearly a finite value to the cross product. Using the hydrostatic relationship for pressure and the observed temperature field we calculate a velocity in the lowest 100 m that increases from about 1 m s^{-1} at PDC-1 to about 6 m s^{-1} at TVR. This tends to overestimate the velocity since it doesn't reflect frictional drag.

In the absence of detailed turbulent momentum flux measurements we made some crude estimates of frictional deceleration due to mechanical turbulence generated by trees and topographic roughness elements in Anderson Creek Valley. The estimates of average frictional deceleration were about half of the average

buoyant accelerations. Application of this factor leads to a typical estimated drainage flow speed of approximately 4 m s^{-1} at TVR. The agreement of this estimate with the observed speeds is quite good and tends to support the idea that we are looking at a buoyancy driven drainage wind in the lowest 100 m in Anderson Valley. Furthermore, the flow is consistent with the temperature and pressure observations.

4. SUMMARY AND CONCLUSIONS

We have analyzed an ensemble of temperature measurements including vertical soundings and surface based measurements from fixed stations and a mobile unit. Although intercalibration was not rigorously conducted between the various systems, each system was calibrated independently and sufficient agreement exists when sensors were co-located to support confidence in the measurements.

Surface measurements at various altitudes in Anderson Creek Valley were as much as 10 centigrade degrees cooler than the temperature in mid valley at the same altitude. The departures were smallest in the early evening and grew during the night. The cold air was quite shallow and if care were taken to make ground-based measurements on promontaries or towers, the gross stability could be characterized by surface temperature observations up the side of the valley.

A more promising analysis of the data is to interpret the thermal gradients from the valley wall to the center in terms of the driving force they provide to the drainage flow. The circulations theorem of Bjerknes provides the link between the stable temperature gradients on sloping ground and the velocity field through pressure-density solenoids. Applying the theorem and accounting for frictional drag effects at the surface, we estimate a drainage domain about 100 m in depth with a speed of 4 m s^{-1} at TVR. Measured profiles at that location show quite good

agreement in a sublayer although larger scale driving forces account for a complex structure of wind speed and direction above the 100-200 m layer.

5. ACKNOWLEDGEMENTS

As a joint experiment, the data base includes contributions from several research groups. The results began to make sense only when we incorporated the tethersonde data of Pacific Northwest Laboratories and the NOAA Wave Propagation Laboratory. The tower data of Lawrence Livermore National Laboratory and Pacific Gas and Electric Company lent further insight. We appreciate the cooperation of all the involved people and organizations. This work was supported by the U. S. Department of Energy, Office of Health and Environmental Research.

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Fig. 1

Fig. 2

Fig. 3

FIGURES 1-3. Temperature profiles for the nights indicated estimated from balloon soundings and surface measurements up the side of the valley.

FIGURE 4. Cross section of potential temperature along the axis of Anderson Creek Valley, July 22, 1979.

FIGURE 5. Cross section of potential temperature along the axis of Anderson Creek Valley, July 24, 1979.

VELOCITY AND TEMPERATURE OSCILLATIONS IN DRAINAGE WINDS

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ABSTRACT

Spectra of the time series of wind speeds and temperatures measured in the outflow area of a large drainage region are presented. Peaks are found in these spectra at frequencies corresponding to periods of oscillation of approximately 1.5 hours. A simple model is given which accounts qualitatively for some of the observed features.

1. INTRODUCTION

In 1968, Tyson reported observations of velocity fluctuations of the mountain wind in the valleys near Pietermaritzburg, South Africa. These fluctuations appeared as surges with periods ranging between 45 and 75 minutes. He presented examples of the variation of logarithmic spectra, $nS(n)$, with height and wind speed, and showed a peak in these spectra associated with the presence of the characteristic surges. Of particular interest was the fact that the peak is found in a region of the spectrum often identified with a "spectral gap" (van der Hoven, 1957; Lumley and Panofsky, 1964) in which the energy density is relatively small.

In this paper we describe some additional observations of peaks in the drainage wind spectrum as well as similar peaks in the temperature spectrum. A simple

model adopted from one proposed by Fleagle (1950) is shown to account qualitatively for some of these observations.

2. DATA COLLECTION AND ANALYSIS

As part of the ASCOT measurement program in 1979, temperature and wind speeds were recorded on a 30m meteorological tower located at the Diamond D Ranch. The tower was equipped with thermistors and 3-component Gill propeller anemometers at heights of 3, 10.1, 22.9 and 29.3 m. Data were collected for various measurement periods from May through September, 1979. During most of these periods, data from all the instruments were collected once every 10 seconds and stored on magnetic tape.

For the spectral analyses, data were averaged in 20 second blocks, and 2048 such data were selected for each run. This covered a period of about 11.5 hours, nominally extending from sunset to sunrise for nighttime runs and from sunrise to sunset for daytime runs. Each series was analyzed with a fast Fourier transfer algorithm after linear detrending. Logarithmic spectral densities were computed for equal, logarithmically spaced frequency bands with 8 bands per decade of frequency.

3. MODEL

Winds in the vicinity of the tower can be expected to show the influence of the Anderson, Putah and Guming Creek drainages as well as several smaller but closer drainage regions. The resultant wind structure is likely to be quite complex, with local flows and pooling of more extended flows interacting in some unknown

fashion. Simple models cannot be expected to predict the details of such wind fields, but they may account for their qualitative features.

Fleagle (1950) proposed a model wherein the forces acting on a cooled layer of air adjacent to the slope produced oscillations in the velocity of this layer. He showed that as the air accelerates down the slope, adiabatic heating results in a reverse pressure gradient which retards the flow. As the air decelerates, radiational cooling increases the pressure gradient and the cycle is repeated. Frictional drag will eventually damp out the oscillations.

Petkovsek and Hocevar (1971) and Manins and Sawford (1979) also developed layer averaged models. Following this approach we may write the governing equations for drainage winds as

$$\frac{\partial V}{\partial t} = - \frac{g D \sin \alpha}{\theta_0} - F_r \quad (1)$$

$$\frac{\partial D}{\partial t} - V \gamma \sin \alpha = \frac{Q}{h} \quad (2)$$

V and D are the bulk velocity and temperature deficit for the cooled layer, respectively, α is the slope angle, F_r is a general frictional force, γ is the ambient potential temperature lapse rate, h is the depth of the layer and Q is the net vertical temperature flux into the cooled layer.

Manins and Sawford (1979) suggest that the most important contribution to F_r arises from entrainment of ambient air into the cooled layer. This may be written in the form

$$F_{r_{\text{entrain}}} = E V^4 / h \quad (3)$$

where E is an entrainment coefficient given by

$$E = \frac{A}{KV^2 - S_1 D h g \cos \alpha / \theta_0} \quad (4)$$

and A , K and S_1 are constants.

Numerical solutions of (1) and (2) using (3) plus a surface drag force of the form $C_D V^2$ leads to damped oscillatory solutions for both V and D . An example of this behavior is shown in Figure 1. For reasonable estimates of the input parameters the period of oscillation is found to be ~ 1.5 to 2 hours.

4. EXPERIMENTAL RESULTS AND DISCUSSION

Wind data from 30 nights were examined. In 26 of these a peak in the logarithmic spectral energy density was found in the frequency range 10^{-4} to 3.3×10^{-4} Hz, corresponding to periods of approximately 1 to 3 hours. The spectra were normalized to their peak values, averaged over the ensemble of 26 nights and normalized to the peak averaged value. The results are shown in Figure 2. There is a well defined peak at a period of about 1.5 hours. While this is longer than the periods found by Tyson it agrees rather well with the values predicted by the simple model outlined above. Moreover, it is well within the frequency range usually associated with the spectral gap.

For comparison purposes 14 days were randomly chosen from the time periods for which the nighttime spectra were determined, and a similar analysis was made. The results are shown in Figure 3. Again, there is a well defined peak in the spectrum, but at a frequency value higher than that of the nighttime peak by almost a factor of 20. The location of this peak agrees well with values generally reported in the literature (e.g., Kaimal et al., 1972).

Figure 4 shows the averaged temperature spectrum for 14 nights selected from the 26 previously used for the wind analysis. A low frequency peak is also found here, and there is close agreement between the frequencies for which the wind and temperature spectra have their maximum values.

The covariances of temperature and wind speed fluctuations were found to be positive in general, although in one case it was negative. The simple theory given previously predicts negative covariances over a fairly wide range of values for the governing parameters, but significant phase differences between the temperature and wind time series are easily produced. It would be desirable to repeat these measurements in an area of simpler topography to see if the failure of the model to reproduce the measured sign of the covariance might be related to the complexities of the drainage region studied here.

ACKNOWLEDGMENT

We wish to thank Dr. Monte Orgill for suggesting this problem for study.

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FIGURE 1. Modeled wind speed.

FIGURE 2. Wind spectra for 26 nights.

FIGURE 3. Wind spectra for 14 days.

FIGURE 4. Temperature spectra for 14 nights.

ANALYSIS OF ACOUSTIC SOUNDER AND TETHERSONDE DATA
OBTAINED FOR THE ATMOSPHERIC STUDIES IN COMPLEX
TERRAIN (ASCOT) PROJECT DURING 1979

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1. INTRODUCTION

During July 1979 and succeeding months a variety of acoustic-sounder data were obtained in the vicinity of an east-facing valley located about 60 km inland from the coast of California in support of the ASCOT program. Within the valley a number of acoustic sounders were deployed to monitor the variations in the height of the nocturnal inversion and boundary layer as shown in Fig. 1. Additionally, a Doppler-sounder unit was located at the mouth of the valley in a field belonging to the Diamond D Ranch. This more-permanent installation provided a climatology of the wind profile to heights of 600 m. A series of tethersonde wind and temperature profiles supplemented acoustic-sounder data. In particular, the Wave Propagation Laboratory operated a tethersonde near a portable acoustic sounder at a site referred to as "Thorne-7" within the valley, at an elevation of about 300 m above the Diamond D Ranch.

During the last night of the July experiment, the acoustic sounder was moved from Thorne-7 to a site near triangulation station TAM on the ridge above the head of Anderson Creek. This site lies about 800 m above the Diamond D Ranch at the mouth of the valley.

FIGURE 1

2. SUMMARY OF DOPPLER DATA

We will first outline results from the Doppler wind measurements at the Diamond D Ranch. These measurements were designed to characterize the vertical profiles of the integrated nocturnal drainage flow out of the valley. This Doppler-acoustic sounder uses a LSI-11 microprocessor and floppy-disk based real-time data acquisition system as described by Neff et al. (1980). Profiles of wind speed and direction are printed every twenty minutes in the form shown in Fig. 2. The data are also recorded on floppy disks for later archival and the generation of time series such as that shown in Fig. 3 and histograms such as those in Fig. 4.

From such analyses of Doppler/monostatic sounder data, we found four classes into which most of our observations fell.

Synoptic/Mesoscale Winds

A westerly wind began regularly in the early afternoon and continued until about 2000 PST each evening with speeds of about 10 m/s at a height of 600 m. Oftentimes, the Doppler wind profiles indicated a clockwise turning of this wind during the night with northeast winds occurring during the early morning hours.

Katabatic Winds

During the period from about 0000-0600 PST, westerly winds were generally weak and/or variable aloft. However, low-level westerly winds occurred in bursts lasting from 20 to 120 minutes and correlated well with changes in the inversion height over the valley. We will refer to these as "intermittent katabatic" flows.

Layering

A much more complex pattern developed during the last week of the ASCOT July 1979 experiment. In these cases, the wind direction varied markedly with

FIGURE 2

FIGURE 3

CDOE WIND SPEED HISTOGRAM

WIND SPEEDS BETWEEN THE HOURS OF 0:00 AND 1:00 (LOCAL TIME).
 READINGS TAKEN EVERY 20 MINUTES AT ALTITUDES FROM 50-600 METERS.

EACH "*" REPRESENTS 15 OCCURRENCE(S).

CDOE WIND DIRECTION HISTOGRAM

WIND DIRECTIONS BETWEEN THE HOURS OF 0:00 AND 1:00 (LOCAL TIME).
 READINGS TAKEN EVERY 20 MINUTES AT ALTITUDES FROM 50-600 METERS.

EACH "*" REPRESENTS 15 OCCURRENCE(S).

Figure 4

height. Our interpretation of the acoustic sounder facsimile and Doppler records shows layers intruding into the valley. Such layers would persist for hours at a time and varied in thickness from 50 to 200 meters deep with wind directions typically between northeast and southeast. Observation of smoke releases near Thorne-7 indicated outflow in thin layers over the valley at times.

Upslope Winds

With the heating of the valley walls in the morning, a thermally-forced upslope flow began around 0800 PST each day at Diamond D Ranch and somewhat earlier at higher elevations in the valley. This thermally-forced flow appears to provide a pumping mechanism such that the inversion over Diamond D Ranch would lower markedly between about 0700 and 1000 PST: the uppermost inversion layer, typically about 600 meters above the surface would lower about 300 meters over this period to meet the eroding convective layer below. This effect was most regular during periods of slack synoptic-scale winds.

3. THORNE-7 MONOSTATIC-SOUNDER AND TETHERSONDE DATA

Although the Doppler wind data enters conventional analyses fairly easily, the interpretation of acoustic backscatter records requires a more subjective evaluation. In particular, the strength of the backscatter is a function of the magnitude of the turbulent velocity fluctuations as well as the vertical potential temperature gradient. In terms of eddy-diffusion coefficients, the backscatter cross section can be written as

$$\sigma_o = 9.0 \times 10^{-6} k^{1/3} K_H^{2/3} \left(\frac{R_f}{1-R_f} \right)^{1/3} \left(\frac{\partial \theta}{\partial z} \right)^{5/3}$$

where k = acoustic wave number, K_H = eddy diffusion coefficient for heat, R_f = flux Richardson number, and θ = the potential temperature. From the second-order closure parameterization of Brost and Wyngaard (1978) the behavior of σ_0 can be predicted (Neff, 1980). In particular, a small change in the shear (8 percent) can lead to an order-of-magnitude change in the backscatter. Mechanisms that accelerate the flow and the variations in density at air-mass boundaries might then be expected to produce large variations in the backscatter profile. This is one reason why acoustic sounder data have proven most useful in the study of internal waves (that can increase both the shear and temperature gradients locally) and the boundary layer (where the no-slip condition at the lower boundary requires a large shear in the presence of static stability).

The fact that the magnitude of the backscatter is several steps removed from knowledge of the mean shear and temperature gradient is restrictive. However, the continuity and high density of the data required make it particularly useful in combination with other crude-profile measurements such as from tether sondes (which suffer from the effects of spatial and temporal variability in the sampling of the flow).

Using acoustic sounder and tether sonde data obtained at our Thorne-7 site in this manner, we found that the inversion/boundary layer divided into three regions as shown in Fig. 5. The lowest layer is characterized by a strong temperature inversion. In this "katabatic" layer the flow is largely gravity-driven. In the "transition" layer, turbulence generated by the katabatic flow mixes cooler fluid from below into the overlying mesoscale inversion layer. In both the transition and mesoscale inversion layers, buoyancy forces are weaker, allowing synoptic and mesoscale pressure gradients as well as buoyancy accelerations from internal waves to become important.

FIGURE 5

Above the katabatic flow, which is only some 10 to 30 meters deep, there is thus a region whose characteristics depend fairly critically on larger-scale meteorological conditions. These results are similar to those reported by Manins and Sawford (1979). This is the region that we have emphasized in our analysis, rather than the shallow surface flow that depends so strongly on local topography.

In the several sections that follow, we will examine some of the principle features revealed by the acoustic sounder and tethered sonde observations. In addition, we will discuss some of the measurements that would be useful in future experiments.

4. SYNOPTIC/MESOSCALE WINDS

A number of authors have discussed the larger-scale wind fields of the Pacific Coast region. In particular, the combination of the eastern Pacific subtropical high and the heat-low of the interior valleys of California lead to a strong onshore flow (the Pacific Coast monsoon) with a superimposed sea breeze. However, because of the complexity of the terrain, the character of the sea breeze is not of the classical variety (Fosberg and Schroeder, 1966). As in the earlier work of Schroeder (1961) and Frenzel (1962), Fig. 3 shows a distinct wind increase in the afternoon and a wind decrease in the late evening after the nocturnal inversion has developed. In most cases the wind speed increases with height to the maximum range of the sounder (600 m). Analyses by Fosberg and Schroeder (1966) showed westerly streamlines over this region in mid-afternoon. In addition, their hygrothermograph recordings at Middletown revealed dew points very close to those measured near the coast. The relative humidity, however, showed a strong diurnal variation because of the heating during the day, particularly over the west-facing slopes in the afternoon.

The boundary-layer echoes from the monostatic sounders in these cases have a generally uniform appearance. The depth of the boundary layer depends roughly on wind speed but lies in a range from 200 to 400 m deep. Its depth does not vary greatly from point to point within the valley. Figure 6 shows a typical backscatter record obtained at the Diamond D Ranch while Fig. 7 shows the variation in wind speed and direction at height 600. The effect of the Coriolis force in this case appears to cause the accelerated flow to rotate clockwise with time, similar (but of opposite sense) to the rotation observed by Manins and Sawford (1979) in Australia. The inertial rate of rotation (360 degrees in a pendulum day of about 18 hours) is also plotted on Fig. 7. The elevated echo layer that forms aloft around 1800 h may reflect the effect of surface cooling as described by Neff (1980) from second-order closure model simulation of acoustic sounder records. Another possibility is that subsidence intensifies an elevated inversion leading to increased scattering. However, examination of weather maps for this period showed no large scale anti-cyclonic circulation. The complexity of the local terrain and the buildup of a large scale inversion might be expected to obscure the observation of such inertial motions. The presence of Cobb Mountain, in particular, would block the flow within the valley for some wind directions. The inversion, which can often be seen (visually, as marked by a haze layer and with acoustic sounders) at ridge level would isolate most measurement systems from observing such a rotation of the wind vector. For this reason, future measurements might include a wind measurement on the top of Cobb Mountain to provide at least some more extensive statistics on the mountain-top winds.

The boundary layer flow that develops in response to such strongly driven circulations, although not purely katabatic in nature, will perhaps be the most successfully modeled flow in the Anderson Creek drainage. The reasons for this are that the thermal forcing is small, the eddy diffusion is large, and mass consistency

DIAMOND D ACOUSTIC SOUNDER

Figure 6

19 SEPTEMBER 1979

FIGURE 7

and overall static stability may be the major constraints on the flow. The event of July 22, 1979, although not of the sea breeze type, is indicative of this type of flow. In such cases the temperature of a parcel descending along the valley walls does not cool significantly due to radiative flux divergence. Nor because of the parcel's short residence time does the turbulent heat flux divergence result in a significant temperature change. Rather it appears that adiabatic heating may be the most important factor: a descent of 400 meters will result in a heating of about 4°C . In the July 22 case, such adiabatic heating of the parcel is sufficient for its temperature to be warmer than the air in the lower part of the valley where the second tracer was released.

It would then appear that such cases of strong westerly winds would be best studied with meteorological measurement systems and tracer releases oriented along the centerline of the valley which runs approximately east-west.

The most important measurements should be:

1. Temperature and wind measurements with tethersondes for an isotropic analysis over the valley as well as the wind structure.
2. Tracers should be released at two heights at a single site (the stronger the winds, the greater the optimal separation) to estimate the efficiency of vertical diffusion.
3. Tracers should be released at more than one elevation to determine the effect of stability on the separation of the flow from different source regions. Temperature mapping will also help here.
4. Sampling should at least occur in combination with fairly complete meteorological measurements. Such sites should include a surface wind-temperature station to provide a continuous record of near-surface conditions, tethersondes, and a monostatic acoustic sounder. Since the tethersondes profile can be significantly distorted by intermittency in the

flow, the acoustic sounder can provide a more detailed evaluation of boundary layer characteristics.

5. INTERMITTENT KATABATIC FLOWS

Intermittent katabatic flows appear to be a significant feature of Anderson Creek Valley. Figure 8 shows such an outflow at Diamond D Ranch while Fig. 9 shows the corresponding record from Thorne-7, some 300 m above the elevation of Diamond D Ranch. A gentle inflow is seen to precurse the outflow event. Figure 10 shows the relation between the volume outflow during these events and the change in the volume of the inversion above the valley. Horizontal homogeneity within the valley is assumed in this calculation. These data suggest that the entire "mesoscale" inversion structure is involved in such events suggesting the presence of internal wave modes or perhaps even seiches that interact with the topography. The mass outflow during such events is directly related to the change in the inversion depth over the valley. Weak inflows observed from the Doppler unit at Diamond D Ranch can serve as a predictor of such outflows some 40 to 80 minutes prior to the event with some 50% reliability based on the analysis of July 1979 data. Cases without an inflow precursor were characterized by stronger large scale westerly flow: typical inflow speeds are about 1 m/s and are thus easily masked. With the analysis of the additional data gathered during 1979, we hope to be able to improve our predictive ability. This would allow closer coordination with tracer releases.

1. Our calculations indicate that the change in the volume under the valley inversion was nearly equal to the outflow volume during intermittent katabatic flows. The inversion depth should therefore be measured on an array over the entire valley with emphasis on the lower half of the valley where pooling or convergence might be expected.

DIAMOND-D

Wind Components Along 290° (m/s) from Acoustic Doppler System

FIGURE 8

FIGURE 9

FIGURE 10

2. An acoustic sounder should be placed some 5 km outside the valley to examine variability in the inversion height should the valley flow be coupled to larger-scale modes.
3. A reduced power lidar (say eyesafe at 1 km) could be used effectively at the mouth of the valley to trace the advection and growth of an oil fog plume (the growth being due to either accumulation in the lower portion of the valley or diffusion). Scanning the plume over ten minute periods should provide adequate resolution of these events that last an hour or more.

6. LAYERING AND INFLOW CASES

These cases occurred towards the end of the July 1979 experiment. Figure 11 shows typical acoustic echoes in such events. The clear echo regions generally identify the inflowing air masses. Northeasterly winds across the top of the inversion appear to generate considerable overturning and mixing. The only precursor suggested by the acoustic sounder data was the appearance of an elevated echo layer at about 500 m (as in Fig. 11) during the latter half of the sea breeze period (1600-2000 PST). Inspection of Doppler data during these events indicated a stronger (8 m/s) low-level flow (200 m) around sunset. One possibility is that the increasing stability (and reduced Froude number), particularly near the mouth of the valley, causes the flow to accelerate there in the lowest layers as it exits the mouth of the valley.

To examine the character of such inflows in more detail, we carried out a short field experiment in July 1980, using three monostatic acoustic sounders located along the axis of the valley. The onset of southeasterly flow into the valley during the night of July 15-16 directly correlated with a lifting inversion at that

DIAMOND D ACOUSTIC SOUNDER

FIGURE 11

time, as revealed in Fig. 12. Prior to the inversion rise at 0020 PST, the acoustic facsimile records showed the inversion base at both Diamond D Ranch and Cal Trans at 200 m. The sounders operated 3.5 km apart at approximately the same elevation. At the beginning of the inversion rise at Diamond D Ranch and Cal Trans, a nocturnal inversion, 200 m in depth, existed at the Thorne-7 site.

Wind speeds and directions obtained from a tether sonde flown at Diamond D Ranch showed a weak inflow (0.6-0.7 m/s) at 0025 PST between 130-170 m. This region corresponds to the clear region below the rising inversion on the facsimile records. An inflow greater than 1 m/s at 0142 PST between 270-333 m again corresponds to the clear region below the rising inversion. By 0148 PST an inflow greater than 1 m/s expands to between 236-268 m while the inversion continues to rise.

For the 60 minute period following the initial increase in inversion height, the inversion rose at a rate of 2.3 m/min at Cal Trans and 2.1 m/min at Diamond D Ranch. Thorne-7 experienced a much steeper rise of 4.4 m/min. The inversion rise at Thorne-7 occurred 52 minutes after the initial rise at Diamond D Ranch. Theoretically, with an initially horizontal inversion over the valley and a rise rate of 2.1 m/min as experienced at Diamond D Ranch, it would take 48 minutes for the inversion to rise at Thorne-7.

Such events appear to depend on the amount of penetration of marine air inland. The approach of an upper-level trough with its associated surface front will normally cause an increase in the depth of the marine layer due to the decrease in the subsidence aloft and convergence ahead of the surface front. Such an event was observed at the ASCOT experiment area during August 1980, with stratus flowing over the ridges southwest of Middletown and marked inflow over the Diamond D Ranch site from the southeast. Prior to the passage of an upper-level trough at 0000Z on the 19th, the marine influence became more pronounced in the ASCOT

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FIGURE 12

study area. Table 1 summarizes data obtained from the Oakland upper-air soundings taken during this period. The August 17th 0000Z sounding shows typical marine inversion heights and strengths observed when the area is under the influence of the Pacific anticyclone at 500 mb. The marine inversion decreases in both height and strength during the next 24 hours. The August 19th 0000Z sounding shows the effects on the depth of the marine layer due to the approach of an upper level trough. The inversion remains weak but increases in height by 500 meters in a 24-hour period. This also corresponds to the period during which the flow in the Middletown area was undergoing the greatest changes. Due to the presence of a surface cold front south of Oakland, difficulties arise in whether to classify the inversion as subsidence or frontal. The August 20th 0000Z sounding again shows typical heights and strengths for a subsidence inversion at the 500 mb anticyclone reestablishes itself over California.

7. UPSLOPE WINDS

With the onset of solar heating, the nocturnal flow reverses direction and generally easterly winds prevail within the valley. The effect on the inversion that

TABLE 1. Data obtained from Oakland soundings.

Date	Time	Inversion Heights	Inversion Temperatures
17 August	0000Z	479-1114 m	15.2°-21.9°C
18 August	0000Z	264-439 m	15.4°-16.0°C
		1710 m	14.6°C
19 August	0000Z	763-911 m	14.2°-15.8°C
		1301 m	16.4°C
20 August	0000Z	316-869 m	15.3°-21.2°C

overlies the Diamond D Ranch is also quite noticeable as can be seen in Fig. 13 where elevated inversions descend to meet the inversion that caps the near-surface convective region.

8. THE BOUNDARY LAYER ON THE RIDGE

As we noted earlier, we moved our portable acoustic sounder to a point near triangulation station TAM on the last night of the July ASCOT experiment. Figure 14 shows the results obtained. Winds on the ridge were westerly at about 10 m/s during the afternoon and early evening. As the winds decreased after 2200 PST, the boundary layer (delineated by the region of strong temperature fluctuations) grew to about 100 m in depth. After 0300 PST on July 28, and as the wind (measured at the Diamond D Ranch turned to northeast) an elevated inversion with large amplitude waves imposed on it developed some 300 m above the top of the ridge. About one hour after the valley side walls received solar insolation, the region of strong returns grew to the height of the inversion. After 1000 PST on the 28th, wind noise obscures the record.

DIAMOND D ACOUSTIC SOUNDER

Figure 13

TAM ACOUSTIC SOUNDER

FIGURE 14

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DIURNAL VARIATION OF HORIZONTAL WIND DIRECTION FLUCTUATIONS
 σ_θ IN COMPLEX TERRAIN AT GEYSERS, CAL.

by

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Abstract

Horizontal diffusion in the surface layer is dependent on the standard deviation of wind direction fluctuations σ_θ . Diurnal variation of this parameter in complex terrain was studied for the July 1979 Geysers, Cal., experiment, using data from a network of 11 short meteorological towers in the 25 km² Anderson Creek watershed. Valley side slopes are roughly 20° and maximum terrain difference is about 1 km.

Values of σ_θ for wind directions sampled for one hour at a height of 10 m are about 35° during the daytime. They slowly decrease to about 20° by 8 to 10 p.m. as stability increases but wind speeds are still relatively high. After 10 p.m. the drainage flow sets in at most stations, with speeds of 1 to 2 m/s, and average σ_θ increases to about 30° during the period 11 p.m. to 6 a.m. In general, highest values of σ_θ at night are associated with lowest values of wind speed and greatest static stability. This enhancement of σ_θ by the terrain suggests that horizontal diffusion at night always conforms to that expected during nearly neutral stabilities. That is, Pasquill class D diffusion applies to the horizontal component all night in complex terrain.

1. Introduction

During July 16-28, 1979, several research groups converged on the Anderson Creek basin in the Geysers geothermal area, California, as part of the Department of Energy-sponsored ASCOT program (Atmospheric Studies in Complex Terrain). The purpose of this field study was to take detailed measurements of drainage winds in complex terrain (Dickerson and Gudickson, 1980). Data from tethersondes, acoustic sounders, meteorological towers, a laser anemometer, neutral balloon releases, pibal releases, and tracer releases are described in the report ASCOT Data July 1979 (Lawrence Livermore Laboratory, 1979).

There is currently great interest in diffusion in complex terrain because of recent EPA regulations requiring the use of diffusion models for assessing environmental impact, and the tendency to site new power plants in mountainous regions of the United States. This report is concerned mainly with methods of estimating the cross wind diffusion parameter σ_y in complex terrain. It was recognized at an AMS workshop (Hanna et al., 1977) that the standard deviation of wind direction fluctuations, σ_θ , should be used to estimate lateral diffusion. Pasquill (1976) provided a simple empirical formula for estimating σ_y from σ_θ :

$$\sigma_y = \sigma_\theta \times f(x) \quad (1)$$

where the dimensionless function $f(x)$ is given in Table I:

Table I
Variation of $f(x) = \sigma_y / \sigma_\theta x$ with distance x (Pasquill, 1976)

$x(\text{km})$	0	0.1	0.2	0.4	1	2	4	10	>10
$f(x)$	1.0	0.8	0.7	0.65	0.6	.5	.4	.33	$.33(10/x)^{1/2}$

The function $f(x)$ is dependent on scales of lateral turbulence and the values in Table I represent averages obtained from several experiments. Thus over complex terrain $f(x)$ may be different from Table I due to the presence of mesoscale eddies.

Panofsky et al. (1978) and Hanna (1980) showed that complex terrain could significantly enhance σ_θ by factors of 1.5 to 5.0 over the value expected from similarity theory based on characteristics of the local ground surface. During neutral conditions, similarity theory predicts that σ_θ is given by the formula:

$$\sigma_\theta \approx \sigma_v/u = 0.8/\ln(z/z_0) \quad (2)$$

where σ_v is the standard deviation of lateral wind fluctuations, u is the wind speed, and z_0 is the local roughness length (within a few hundred meters of the tower). This formula is based on the assumption that conditions are horizontally homogeneous, which must be tempered in complex terrain to the assumption that conditions are locally horizontally homogeneous. Presumably the enhancement of σ_θ in the experiments mentioned above is due to terrain-generated wind direction meanders which directly affect the low frequency portion of σ_θ . The purpose of the analysis of Geysers σ_θ reported here is to determine the characteristics of σ_θ at this complex site, and make recommendations regarding the choice of stability classes and σ_y in diffusion models for complex terrain.

2. Data

Measurements of σ_θ from 11 meteorological towers in the Anderson Creek basin of the Geysers geothermal area were analyzed. We used data

from seven 10 m towers (Numbers 1-7) operated by Lawrence Livermore Laboratory (LLL), one 30 m tower (Diamond D) operated by Pacific Northwest Laboratory (PNL), and three 2 m towers (102, 104, 105) operated by Los Alamos Scientific Laboratory (LASL). Standard research-grade cup anemometers and vanes were used on all towers. Data from the 10 m level of the PNL Diamond D tower were used, so that the instrument heights would agree as much as possible. Sampling period for σ_θ data reported by the LLL and the PNL towers was one hour. Values of σ_θ from the three LASL towers were estimated to be 1/6 of the wind direction range measured over an hour. Locations of the towers are shown on the topographic map in Figure 1. In general the LLL towers were well-exposed, the LASL towers were located in canyon bottoms, and the PNL Diamond D tower was in a flat open field. The terrain at all sites but Diamond D was rugged, with nearby rock outcrops and scattered trees and shrubs.

Five days during the July 1979 intensive study period were used in this analysis (noon to noon on 18-19, 22-23, 24-25, 26-27 July). These days were characterized by similar meteorological conditions, with high and low temperatures of about 35 C and 15 C in the valleys. Afternoon winds were westerly (~5 m/s) because of the influence of the dominant westerly synoptic flow and the sea breeze, and morning winds were easterly (~2 m/s) in response to the sun's rays striking the east-facing Anderson Creek valley. On sheltered slopes and in creek canyons, drainage winds (1-2 m/s, with directions perpendicular to terrain contours) dominated the nighttime flow. These characteristics can be seen by examining the hourly wind vector diagram from station 2 in Figure 2 for the five days of study.

3. Analysis of σ_θ Measurements

The average diurnal variation of σ_θ was calculated for each of the 11 stations for the five days under investigation. The highest, median, and lowest of the 11 σ_θ values at each hour are plotted in Figure 3. Average daily σ_θ (one hour sampling time) on the figure is about 28° . With light easterly winds and strong solar heating in the morning, median σ_θ increases to its daily maximum of 44° at 1100. After 1100, the westerly wind begins and steadily increases throughout the afternoon, causing the mechanical component of turbulence to increase relative to the buoyant component. Stabilities trend more and more toward neutrality, and σ_θ slowly decreases throughout the afternoon, reaching a minimum of 19° at 2000. At this time winds are strong at the exposed stations.

Stability increases after 2000 as the wind dies down at the exposed stations and surface based inversions are formed under clear skies. Similarity theory predicts that σ_θ should decrease under these conditions:

$$\sigma_\theta (\text{in radians}) \approx 0.8/(\ln(z/z_0) + 4.5z/L) \quad (3)$$

where L is the Monin-Obukhov length. This equation is valid for stable, neutral, and slightly unstable conditions. A good guess for the roughness length over this terrain would be 1 m. Table II gives predicted values of σ_θ from equation (3) for various values of L :

Table II
Predicted σ_θ for Height 10 m from Equation (3) for Various Values of L

L (m)	-100	∞	100	20
Stability Class	C	D	E	F
σ_θ (degrees)	25	20	17	6.8

The relation between stability class and Monin-Obukhov length was taken from Golder (1972). However, Figure 3 shows that the observations do not follow the trend of σ_θ with stability predicted in Table II. Instead, σ_θ holds a relatively high value (20 to 30°) all night. Terrain-induced mesoscale eddies are causing wind direction fluctuations that would be absent over smooth terrain.

High values of σ_θ can also occur during stable, light wind conditions over flat terrain. Van der Hoven (1976) documented experiments at five sites during very light wind ($0.2 < u < 1.9 \text{ m/s}$) nighttime conditions (so-called class G), when σ_θ was measured to range from 10° to 72°. Enhancement of σ_θ in these conditions is caused by wind direction meander induced by minor perturbations. However, these are extreme, near-calm conditions. At the Geysers site, σ_θ was enhanced not just for these extreme conditions, but for all stable nighttime conditions (classes E, F, and G).

The correlation between σ_θ and wind speed evident in Figures 2 and 3 suggests that wind speed u may possibly be used to estimate σ_θ during the night (2100 to 0600). Nighttime hourly average σ_θ was plotted against u for the 11 stations under investigation, and a typical example of the resulting graphs is given in Figure 4. All reporting stations produced similar curves, with no more than a factor of two difference in the average σ_θ at a given wind speed among stations. The data in the figure are from station 2, which is a relatively exposed site about halfway down the ridge. There is a clear dependence of σ_θ on wind speed u , in the opposite sense of that predicted by similarity theory for stable conditions. At wind speeds greater than about 5 m/s, when the generation of turbulence by mechanical effects is high, nearly neutral conditions prevail and observed σ_θ averages about 8°. The discrepancy between the observed value (8°) and the predicted value (20°) in Table I is probably due to problems in estimating the

effective roughness and the height of the instrument above the local roughness elements in such variable terrain and vegetation.

Figure 4 shows that σ_θ increases to about 50 to 100° at low wind speeds, but there is much scatter. In no case is hourly σ_θ less than 20° for wind speeds less than 2 m/s. It is not possible to know whether a pollutant plume released under light-wind conditions at night will be relatively thin ($\sigma_\theta \cong 20^\circ$) or severely distorted ($\sigma_\theta \cong 100^\circ$). Similar scatter occurred in the σ_θ versus u diagrams for the other stations. This is a strong argument for monitoring σ_θ on site at all times for use in diffusion models.

The simple relationship

$$\sigma_\theta u \cong \sigma_v = \text{constant} \quad (4)$$

provides a surprisingly good fit to the data in Figure 4, even though there is much scatter in σ_θ at low wind speeds. The solid line represents $\sigma_v = 1.0$ m/s. Apparently nighttime flow in the Geysers complex terrain adjusts itself so that the turbulent energy, proportional to σ_v^2 , is constant independent of wind speed or stability. Another implication of this interesting observation is that the product $\sigma_y u$ which appears in the denominator of the Gaussian plume equation equals a constant, $\sigma_v x$, at any given distance downwind at night in complex terrain.

These data show clearly that σ_θ is consistently enhanced during nighttime stable conditions with light to moderate wind speeds in the Geysers area. Average σ_θ is about 25° (.44 radians) for the night period at a height of 10 m for 11 stations located in rough terrain.

Consequently standard applications of the Gaussian plume model using Pasquill σ_y curves for stability classes E and F would overestimate plume centerline concentrations on the average in this region. If no σ_θ data are available, it is reasonable to use neutral (stability class D) Pasquill σ_y curves for all nighttime conditions in complex terrain. Crosswind diffusion calculations should be closely watched in case the plume width approaches the valley width. These assumptions are presently being made in the Rough Terrain Diffusion Model developed by Egan (1980), and are consistent with the few observations made in complex terrain reported by Egan (1975). Of course, if σ_θ is measured, it should be used directly to estimate σ_y through the relation $\sigma_y = \sigma_\theta x f(x)$.

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Figure 1. Topographic map of Anderson Creek basin, located in Geysers Geothermal Area, California, showing meteorological tower locations. LLL (1-7), LASL (102-104), PNL (Diamond D).

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Figure 2. Hourly average wind vectors for five days at station 2.

Figure 3. Hourly average σ_θ values at a given time of day were averaged together for five days for each of 11 wind stations. The highest, median, and lowest of the 11 σ_θ values at each hour are plotted on this graph.

Figure 4. σ_θ versus wind speed u for nighttime (2100-0600) data from five nights for station 2. The solid line is the equation $\sigma_\theta u = 1.0$ m/s.

APPLICATION OF A PHOTODIODE-ARRAY OPTICAL TURBULENCE SENSOR TO WIND STUDIES IN COMPLEX TERRAIN

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ABSTRACT

A digital photodiode-array optical turbulence sensor was used to gather data simultaneously with analog optical anemometer measurements during the July 1979 ASCOT experiment. This system provided useful information regarding the uniformity of optical turbulence used by the optical anemometer to derive cross-path wind speeds. Wind speeds derived from digital analysis of the photodiode-array intensities also provided an independent measure of the cross-path wind speed. Close agreement was found between these two measures of the wind. The optical turbulence along the 625 meter light path was found to be predictably uniform when averaged over a 10 minute time period at night during drainage wind conditions. At sunrise one side of the valley was heated first and a nonuniform optical turbulence pattern resulted.

INTRODUCTION

The two companion papers to this addendum paper describe results using a laser space-averaging optical anemometer during the July, 1979 ASCOT experiment in the complex terrain Geyser's region.^{1,2} Since this instrument depends on spacial uniformity of optical turbulence to represent a spacially averaged wind and index of refraction structure parameter C_n , it is important to understand when and over what averaging times this assumption is valid in complex terrain. To do this, a simultaneous measurement of optical turbulence along a closely parallel path was made using a digital photodiode-array incoherent light system from selected periods during the drainage wind experimental period. Lawrence and Strohbehn³ have shown how the correlation function for irradiance fluctuations can be used to derive information regarding the uniformity of optical turbulence along a light path. This function along with an independent measurement of the cross-path wind speed was determined from the digitized signals on the photodiode-array elements. This digital technique has several advantages over analog wind systems in that many different analysis techniques can be used on the same set of data. A relative disadvantage is that since each scan represented two seconds of data digitized at 500 Hz, for 16 separate channels, 80 floppy disks are required to store 8 hours of 1 minute data. However, for short intensive experiment periods like the Geyser's study, the results showed the two systems complimented each other quite well.

The results of this comparison showed the following information regarding uniformity of optical turbulence and wind speeds during drainage and morning upslope winds across Anderson Creek:

1. During drainage conditions the irradiance correlation function showed a normal but variable decrease with cell separation associated

with random inhomogeneities in optical turbulence over a two second time period.

2. Ten minute averaging of 10 two second samples removed almost all this variability.
3. However, after sunrise the receiver side of the light path was preferentially heated and the correlation function showed a sudden change (a faster dropoff with cell separation).
4. Wind speeds derived by both the analog laser and incoherent digital systems compared well (within 0.8 m/s) during the whole period even though the spacial filtering for the two systems causes each to have a somewhat different wind weighting function along the path.

The combination of the digital and analog systems allowed ten days of continuous wind and optical turbulence data from the analog system with periodic intensive documentation of the optical turbulence from the digital system.

RESULTS

The experimental set-up and results from the analog laser cross-wind anemometer has been described in the two companion papers from the Gettysburg ASCOT meeting. Figure 1 shows the experimental arrangement for the digital photodiode-array optical turbulence system. The light path was parallel to the laser system (separated by about 1.5 meters) over the same 625 meter path length. The transmitter consisted of a 15 cm mirror with a small white light source at the focus. The existing light beam size was stopped down to a diameter of 11.5 cm. The receiving optics consisted of a 30.5 cm reflecting telescope with a transfer lens, cylindrical lens and a linear 16 element photodiode-array inside the effective focus

of the telescope. The photodiode-array was aligned so that the center half of the focused light illuminated all 16 cells. This was done to limit diffraction effects and small nonlinearities in the photocell responses. The cylindrical lens was used to expand the effective vertical extent of the cells to limit partial coherence effects. The photodiode-array was clocked at 500 Hz and the analog signal from each cell was digitized for 1024 scans (about 2 seconds). These signals were then sampled by a microcomputer and stored on floppy disk media. These data were then analyzed after the experiment. Data were taken at selected periods during each of the 5 nighttime drainage intensive experiment periods. In this paper, we will only describe the results from the data taken on 19 July 1979. Alignment problems on subsequent nights made much of this later data suspect. Also, this was the period chosen for comparison of acoustic and optical turbulence in one of the companion papers.

The theory of how optical turbulence is used to derive spatially averaged cross-wind speeds have been well described by others.^{4,5} The following assumptions about the statistical properties of the optical and atmospheric turbulence are necessary for the derived wind speed to be correct and represent a spatial average with a weighting function peaked at the center of the path and smoothly decaying to zero at each end:

1. The absence of optical turbulence saturation effects (this condition was met throughout the experiment).
2. Kolmogorov inertial subrange atmospheric turbulence statistics for the scales of turbulence affecting the optical measurements (i.e., scales greater than 20 millimeters and less than 15 meters).
3. Homogeneous isotropic optical turbulence along symmetric parts of the path (Gaussian log amplitude fluctuations and therefore log-normal intensity fluctuations are also assumed).

It is this third assumption which is the most difficult to justify in complex terrain even over symmetrically varying topography. Lawrence and Strohbehn³ have shown how the irradiance correlation function can be used to demonstrate non-symmetrical optical turbulence along a light path. The principal application of the digital-photodiode array system was to monitor this function.

Figure 2 shows two examples of the irradiance correlation function on the morning of July 19. These two samples were chosen to show the effect of nonuniform heating of the rising sun on the receiver side of the optical path at 0635 PST when compared to a typical correlation function during the drainage wind night (0423 PST). The values for the cell separations shown in Figure 2 represent the effective size of the distance between cells through the telescope optical system. For a point light source and point receivers, the correlation function should reach zero before one Fresnel zone⁴ (~ 2 cm). However, the extended light source size and detector size has the effect of extending the significant correlation sizes to larger values so that the function at 0423 PST represents the more symmetric optical turbulence distribution of the two situations shown in Figure 2. A parameter r_c is defined to be that value of cell separation at which the correlation drops to 33%. This parameter is used to show how the irradiance correlation function varied throughout the night and morning. Figure 3 plots r_c versus time again for 19 July 1979. The dots represent individual 2 second samples and the open circles and line represent 10 minute averages. This figure shows that over a 10 minute average most of the inhomogeneity of the path averaged optical turbulence is smoothed out. Figure 3 also shows the rapid change in r_c after the sun began preferentially heating the receiver side of the optical path (about 0530 PST) and return to a more symmetric pattern after about 0730 PST. Figure 4 shows these two periods designated on plots of acoustic sounder return strength (function of C_T the

temperature structure parameter) and the standard deviation of the log intensity of the laser σ_I (function of C_N the index of refraction structure parameter). This figure shows that though the rising sun disrupts the horizontal optical turbulence homogeneity it stabilizes the vertical temperature differences decreasing optical and acoustic turbulence.

The digital photodiode-array system was also used to provide an independent measure of the cross-path wind speed. By summing the left and right halves of the array independently, the correlation function can be used to interpret the wind speed. The weighting function for an extended incoherent wind speed determination from the slope of the correlation at zero time lag is more peaked than the analogous weighting function using a laser.⁶ Therefore, a strong relationship between the two techniques is another indication that the weighting function is symmetric in complex terrain when a proper averaging time is chosen. Figure 5 shows a scatter plot with 10 and 30 minute averaging times between the two techniques. The rising sun generated the eastward slope winds (shown as negative values) and there is somewhat more scatter in these results than the westward drainage wind region (positive wind speeds). The transition between drainage and upslope wind occurred at different times at the two ends of the path. This is shown in Table 1. Table 1 shows a 10 minute average time history of the parameters describing cross path wind speed and optical turbulence. It is interesting to note that the transition from downslope to upslope wind occurs simultaneously measured with the two optical anemometer systems. This transition is between the time this 180° wind direction change occurred on the heated side of the path (Aminoil pipeyard) and the shadowed side (Thorn 7). The optical turbulence parameters anticipate this transition.

Our future plans include calibrating the photodiode-array system on an anemometer test range. This will allow us to include other properties of the

correlation function such as the width of the peak, what frequency separation balances the functions, the time delay to maximum, the spacial filter spectral peak and the optical turbulence parameters σ_I and r_c in the wind speed and weighting function determination. This system will be used again in the September 1980 ASCOT Geyser's area experiment, and results will be compared with 7 other optical anemometer measurements.

CONCLUSIONS

The irradiance correlation function and the wind speed determined with a digital photodiode-array optical system indicates that the path averaged optical turbulence was symmetric for the particular optical path used in the July 1979 ASCOT experiment when time averages greater than 10 minutes are used. During the morning, and presumably in the evening, the optical turbulence predominates on the heated side of the light path making the wind weighting function asymmetric. A comparison of cross-wind measurements in the morning show that the transition from downslope drainage winds to upslope winds occurred first on the heated side of the optical path, then on the two optical anemometer readings and finally 30 minutes later on the shadowed side of the path as the sun began to heat it.

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TABLE 1. Ten and Thirty Minute Averages of Wind and Optical Turbulence Parameters on the Morning of July 19 in the Geyser's Region.

Time (PST)	Laser Anemometer Cross-Path Wind Speed (m/s) (10 min avg)	Thorn 7 Cross-Path Wind Speed Component (m/s) (10 min avg)	Aminoil Pipeyard Cross-Path Wind Speed Component (m/s) (30 min avg)	Standard Deviation of Log of Laser Light Inten- sity (σ_p) (10 min avg)	Slope of Time-lagged Correlation Function from from Photo- diode-Array System (10 min avg)	Spatial Correlation Distance from Photodiode Array System (cm) r_c (10 min avg)
0420	2.8	1.27		0.11	.075	2.21
0430	2.3	0.19	0.32	0.11	.064	2.90
0440	2.1	0.22		0.12	.054	2.05
0450	1.9	0.00		0.11	.052	2.34
0500	1.7	-0.10	0.19	0.10	.053	2.40
0510	1.8	0.59		0.09	.038	2.33
0520	1.8	1.50		0.09	.053	2.44
0530	1.7	0.92	0.28	0.09	.067	2.46
0540	1.8	0.91		0.08	.029	1.74
0550	0.6	0.73		0.09	.027	1.70
0600	0.5	0.63	0.11	0.06	.008	1.19
0610	0.5	0.42		0.04	.020	1.16
0620	0.7	0.85		0.05	.029	1.14
0630	0.9	0.91	-0.35	0.03	.021	1.18
0640	-0.1	0.19		0.05	-.012	1.19
0650	-0.4	-0.05		0.04	-.013	1.09
0700	-0.7	-0.90	-0.78	0.05	-.014	0.99
0710	-0.8	-0.66		0.05	-.013	1.27
0720	-0.8	-0.57		0.08	-.022	1.18
0730	-0.9	-0.43	-0.76	0.09	-.031	1.15
0740	-0.7	-0.06		0.10	-.026	3.42
0750	-0.7	-0.18		0.10	-.015	3.47

FIGURE 1. Block diagram of photodiode-array wind and optical turbulence system.

FIGURE 2. Plots of typical irradiance correlation functions for nighttime drainage (0423 PST) and after sunrise (0635 PST) versus equivalent cell separation for 2 seconds of optical turbulence data sampled at 500 Hz.

FIGURE 3. Plot of the time history of the correlation distance parameter r_c (distance correlation goes below 33%) during the morning of 19 July 1979.

FIGURE 4. Nighttime drainage ($r_c \approx 2.25$ cm) and heated slope inhomogeneity of optical turbulence ($r_c \approx 1.25$ cm) periods separated from plots of optical turbulence parameter σ_1 and acoustic turbulence (as a function of C_T) versus time for the morning of 19 July 1979.

FIGURE 5. Scatter plot of cross-path wind speeds determined using digital photodiode-array incoherent light system versus wind speeds determined from analog laser system for 10 and 30 minute averaging times.

SPACIALLY-AVERAGED AND POINT MEASUREMENTS OF WIND VARIABILITY IN THE GEYSER'S AREA

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INTRODUCTION

This paper describes the results of a comparison of wind measurements made with conventional cup-vane tower mounted anemometers and optical space-averaged anemometer techniques. The results described cover the period from 7/17/79 to 7/27/79 during the intensive ASCOT experiment in the Geyser's region. The optical anemometer light path is shown in Fig. 1 on a map showing the topography of the Geyser's region and the location of the LLL tower network. The average height of the laser beam above terrain was about 30 meters. Most of the optical anemometer wind data was obtained using a laser beam system described in detail by Lawrence, et al.¹ Some measurements were also made along the same path using a white light photodiode array system developed at LLL. Optical anemometry has several advantages over conventional anemometry in complex terrain studies. These advantages include the following:

1. For nighttime drainage wind studies it was possible to find a path across Anderson Creek in the drainage wind and insensitive to local effects such as side drainage at each end of the path.

2. The spacially-averaged wind measured by the optical anemometer can be more useful to computer modelers who ordinarily must assume wind measured at a point is representative of the spacial averaged winds over grid cells which typically vary in size from 200 m to a kilometer.
3. Since the optical technique involves no moving parts, it is capable of responding to very low winds often associated with nighttime conditions.

The following are some of the disadvantages associated with optical anemometry:

1. Since the system is double ended, two suitable sights must be found for instrument placement.
2. The instrument responds only to the perpendicular component of the wind, so at least two paths would have to be found to determine the average wind direction.
3. Since the wind weighting function depends on the distribution of optical turbulence along the path, measurements need to be time-averaged to truly represent the spacially-averaged wind associated with uniform optical turbulence.

The last point amplifies the difficulty in determining turbulence characteristics of the wind from spacially averaged winds. This difficulty will be described in more detail later in this paper.

The results which follow will be divided into two parts. The first part shows how the point and spacially-averaged winds were statistically related in the Geyser's area. The second part shows the relationship of turbulence parameters in the region. The major results of the statistical comparison show the following:

1. Average winds from the optical anemometer agreed very closely (within 10%) with the path perpendicular winds determined from the cup-vane tower system at one end of the path (Thorn 7). The overall correlation

coefficient was 0.8. However, many times during the experiment the determined wind speeds differed by 2 to 3 m/s. These are times when numerical modelers may find the spacially-averaged winds advantageous.

2. The transition from drainage to upslope winds in the morning occurred first on the side of the light path heated by the sun, then the transition appeared in the optical anemometer data, and finally the shaded side (Thorn 7) responded, all over a period of approximately 30 minutes.

3. Autospectral analysis showed similar spectra for the two types of measurements for frequencies lower than once per four hours.

4. At frequencies higher than once per four hours, the spectra for the point measurement began to flatten while the spacially-averaged measurement continued to decrease from the diurnal peak. This flattening in the point measurement appears related to a spectral gap in the autospectra of the cosine component of the wind direction at about once per four hours.

5. Both spectra showed a pronounced peak at about once per two hours which is apparently due to a regular oscillation in the drainage wind.

In the second part of the results, the variability of wind speed measured by conventional and optical techniques and wind direction changes measured with the LLL tower network are compared for information relevant to atmospheric turbulence parameters. The principle results of these comparisons can be summarized as follows:

1. Estimates of time-averaged truncated energy dissipation are about a factor of two lower for 10 minute averaged winds for the optical system compared to point measurements over a one hour period with spacial averaging.

2. The highest values for the estimated energy dissipation and standard deviation of wind direction (σ_θ) were found during the thunderstorm period of the experiment.
3. Very little correlation was found between σ_θ values at the various stations in the tower network and no dependence with station elevation was obvious in the data.
4. Regression relationships between the various stations worked best during the storm period and for daytime values and worst during nighttime drainage wind and morning upslope conditions.

The next section of this paper details the results summarized in the last two paragraphs.

RESULTS

The first part of this section describes the results of statistical comparisons of the optical anemometer and winds measured at one end of the laser path at Thorn 7. Figure 2 shows the topographic relief below the laser beam (vertical scale exaggerated), the location of the acoustic sounder, and Anderson Creek. The average height of the laser beam was 32 meters above terrain (well within model predictions² and observations³ of the height of the drainage layer). Figure 3 shows the variation of the path-perpendicular component of the wind, over the ten day experiment period, measured with the optical anemometer and the conventional anemometer located at Thorn 7 (both 10 minute averages). Both measurements show the same daily pattern interrupted for two days during a thunderstorm event. The pattern is seen to be much smoother for the spacially-averaged optical technique than the point measurement. These differences are more obvious in Fig. 4. This figure shows a dot plot of 400 values of cross-wind selected from the data

when the regular daily pattern was observed. Though a good correlation is observed (.79) between the two measurements, there are times when the values differ by over 2 m/s. These would be times which numerical models would be wise to either time-average or use the spacially-averaged value. Non-uniform optical turbulence may also, at times, contribute to this discrepancy. At times during the experiment the intensity correlation function was measured with the white light photodiode array instrument.⁴ This technique can be used to select periods and averaging times which demonstrate properties of uniform optical turbulence.⁵

Figure 5 shows the correlation function for the two types of wind measurement and between the optical anemometer winds and σ_θ measured at Thorn 7. These correlation functions were determined for 10 minute data averages encompassing the entire 10 day data set. The correlation function of the Thorn 7 cross wind component and the optical anemometer wind speed shows that changes tend to occur about 10 minutes sooner for the laser wind measurement. Part of this is due to the fact that in the morning the Aminoil Pipeyard side of the path is heated by the sun first. A weather station measurement at this side showed a transition from drainage to upslope winds thirty minutes before the transition occurred at the Thorn 7 side. The optical anemometer showed this transition at an intermediate time. The correlation function for the spacially-average wind and σ_θ shows a negative correlation consistent with high wind low σ_θ relationship described by Hanna.⁶ Table 1 summarizes the statistical relationships between the variables measured by the laser anemometer and at Thorn 7 during drainage wind periods (i.e., excluding the days during and soon after the thunderstorm). These relationships show the following:

1. Essentially the same average winds measured optically and at Thorn 7 (cross-beam component).
2. A significant (F-test $< .01$) positive correlation 0.8.

3. A relatively high correlation between optical turbulence (SIGI) and wind speed at night.
4. At no time did the optical system saturate (i.e., $SIGI \geq 0.3$).
5. The standard deviation of wind direction (SIGT) showed negative correlation with wind speed.

Regular features of time variability were investigated further using autospectral analysis techniques.

Figure 6 shows a comparison of the autospectral densities of the optical anemometer and the Thorn 7 cross-path wind speeds. These data were smoothed with a 30 point (5 hour) triangular filter. Both data show similar spectral density at frequencies less than once per four hours. At frequencies higher than once per four hours both spectra show a relatively strong peak at once per two hours. This is consistent with spectral analyses of the Diamond D tower data.⁷ The slope of the spectral density of the point measurement begins to flatten compared to the space averaged measurement at frequencies higher than once per four hours. To determine whether this flattening was due to wind speed or wind direction changes or both, their autospectral densities are compared in Fig. 7. The filter applied to these data was shorter (5 points), thus the greater variability in the spectra. This figure shows that the spectra for wind speed is relatively flat while wind direction shows a spectral gap at about once per four hours. The two hour peak still exists but is spread out more to other frequencies. Interpretation of this behavior is helped by comparison of these spectra to analogous spectra taken at a point and across a windy pass east of Lawrence Livermore Laboratory⁸ (Fig. 8). These spectra actually show an increase in the point spectra (weather station) with frequencies greater than once per four hours while the laser anemometer spectra continue to decrease. Again, this is seen to be mostly due to wind direction changes. A tentative explanation for this behavior would be that since it has been shown that

complex terrain increases sigma theta at high frequencies this energy must come from somewhere and a spectral gap at lower frequencies like once per four hours might be expected in the component of wind direction at a point. Apparently the exchange of wind speed and direction averaged over larger spacial scales associated with the optical anemometer do not show this effect of the complex terrain. These results imply that point measurements should be averaged for one to two hours to better represent the wind variability over spacial averages of 600 m for 1 km in complex terrain.

Since the major focus of the July ASCOT experiment was nighttime drainage winds, an attempt was made to do spectral analysis on nighttime drainage winds only. This was done by constructing a data set consisting of 8 hour nighttime segments excluding the thunderstorm period. Figure 9 shows the spectra for the space-averaged and point measurements for these reconstructed nighttime drainage wind data. These can be compared to the spectrum of space averaged wind at night during the thunderstorm period when drainage wind activity was disrupted (Fig. 10). The shapes are obviously quite different. Again, the point and space-averaged spectra differ at higher frequencies. All the spectra show evidence of an once per two hour peak. The nondrainage spectra show the dominance of the 8 hour segmenting of the data while the drainage data show less of an effect. Statistics warn us against drawing conclusions based on individual peaks, however the strong peak at about once per 27 minutes in both spectra in Fig. 8 which does not appear in Fig. 10 should be looked for in future experiments along with the once per two hour frequency.

The following section of this report deals with the implication of the measurements relative to turbulence in complex terrain. Since the bulk of turbulent energy is found at very high frequencies, conclusions drawn from five and ten minute averaged and 625 m space-averaged wind variations are very weakly related

to "real" turbulence. An effort is made to try to quantify what is lost in the turbulent energy spectrum by spacially averaging over 625 m. Later in this section, a relationship is derived where the spacially averaged heat flux is estimated from concurrent optical turbulence measurements assuming that an estimate of energy dissipation can be made from the variation in spacially averaged winds. It should be emphasized at this point that the possibility of relating energy dissipation to such large temporal and spacial scale measurements is highly speculative.

It can be shown that assuming homogeneous isotropic turbulence, the energy dissipation ϵ is related to the instantaneous standard deviation of the wind speed component ϵ_u through the correlation function r_t at time t and the mean wind speed u by the equation⁹

$$\sigma_u^2(1-r_t) \propto (\epsilon u t)^{2/3} \quad (1)$$

Since instantaneous measurements of the wind speed were not available, we create a function ϵ' from the standard deviation and autocorrelation of 10 minute averages for one hour and call it a truncated time-averaged estimate of the energy dissipation ϵ . Figure 11 shows the autocorrelation function for 10 minute averaged values of the optical anemometer and cup anemometer at Thorn 7. The autocorrelation after 1 hour is approximately 0.75 for both functions. This is much higher than would be the case for instantaneous measurements of the wind speed. Figure 12 shows corresponding plots of ϵ' for the spacially averaged and point measurements. This figure shows that this parameter is dominated by the thunderstorm period of measurement. ϵ' is greater for the point measurement than the spacially averaged measurement on the average by a factor of 2.4. The absolute value for ϵ' is about 4 orders of magnitude lower than typical measurements of ϵ which implies that with an $r_t \approx 0$, an estimate of σ_u would be a factor of about

5 to low for 10 minute averaged data compared to instantaneous measurements for an hour. The difficulty associated with applying equation 1, with the assumption of isotropic homogeneous turbulence, in complex terrain also compounds the difficulty in relating ϵ' to ϵ . Figure 13 shows that the thunderstorm also dominated 5 minute means of 1 second values of the standard deviation of wind direction (σ_θ) at Thorn 7 with a small positive correlation to ϵ' measured at Thorn 7.

In the next ASCOT experiment we hope to do a better job in trying to relate the spacially averaged wind variations to the real energy dissipation measured with a doppler acoustic sounder. Our motivation for this effort is to estimate the spacially averaged heat flux Q and optical turbulence at night through the relationship¹⁰

$$Q \frac{d\theta}{dz} = C_T^2 \epsilon^{1/3} \quad (2)$$

where θ is the temperature

z is the height

C_T^2 is the temperature structure parameter.

For cases where humidity variation can be neglected C_T^2 is related to the index of refraction structure parameter C_n^2 which in turn is related the standard deviation of the log of the laser light intensity for uniform optical turbulence.

For the present, the best instrumental measure of turbulence in the Anderson Creek, Anderson Ridge area comes from the σ_θ data from the LLL tower network. A time history of σ_θ measured at Thorn 7 is shown in Fig. 13. The autospectral densities for σ_θ at Thorn 7 are shown in Fig. 14. This plot shows a spectral peak at twice per day which is even higher than the diurnal peak at once per day. This differs from σ_θ variation in homogeneous terrain and is consistent with two peaks in σ_θ per day found by Hanna⁶ in a combination of these data and

other tower data in the region. One peak is associated with morning light and variable winds and the other with light drainage winds.

Multiple linear regression analysis was then run for the seven stations in the tower network to look for statistical relationships between the station values for σ_θ . Table 2 summarizes the simple statistical relationships between all the stations derived from the total 10 day data set (5 minute means). The interstation correlations are seen to be quite low (about 0.2) and no consistent variation in average σ_θ with station altitude is evident. Table 3 summarizes the results of the regression analysis. The data used in this analysis were 400 samples from consecutive periods including the thunderstorm period. The stations chosen as independent variables were the T.V. Repeater, Thorn 7, and Anderson Ridge. The variance explained using the other stations as independent variables with an F-test of 0.05 shows that the relationship between the stations was best during or soon after the thunderstorm period, and consistently higher for daytime data than nighttime data.

CONCLUSIONS

Comparison of wind speeds and direction data at a point in complex terrain with spacially-averaged wind speeds using optical techniques provides valuable insight into the spacial-temporal influence of topography on wind variability. Since numerical models to be run on the Geyser's data set will have difficulty getting spacial resolution less than 600 m, these comparisons help in two ways. First, the spacial average wind component can be used directly in the grid cell including the laser path. Secondly, this temporal spectral comparison shows that other wind measurements at points in the region should be averaged for at least one to two hours to better represent the wind variability over a grid cell. Much less success

was found in attempting to relate the point and spacially averaged wind variability to atmospheric turbulence parameters.

These results lead to the following recommendations to improve the September 1980 ASCOT experiment and/or other future experiments in complex terrain:

1. More optical anemometer paths are planned and should help quantify other drainage wind sources in the Geyser's region.
2. The intensity correlation function should also be monitored at least along one laser path to provide information on when or over what time average optical turbulence along the path approaches uniformity.
3. If two paths can be found with one end touching the other displaced by about 30° , where the underlying terrain is similar, this measurement would provide a measure of the mean wind direction. If these path measurements could be maintained over a period of a week or more, needed information relating spectral changes of direction and speed would be obtained to better understand the time variability of point and spacially-averaged winds.
4. Separate measures of turbulence parameters with high frequency response measurements from either doppler acoustic or tower instrumentation or both would be helpful in relating optical turbulence at night to the heat flux through the energy dissipation. Also information on what time scales and space scales (if possible) homogeneous isotropic turbulence assumptions can be made in complex terrain (if any).

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TABLE 1. Statistical relationships between spatially averaged and point measurements in the Geysers region for selected drainage periods (400 data points for 10 minute averages).

Variable Description (Name)	Periods*	Statistics		Value		Correlation	
		Mean (units)	Standard Deviation	Minimum	Maximum	Maximum Absolute	Variable Name
Optical Anemometer	Day	2.09 (m/s)	1.93	-1.9	6.6	.62	ULCOMP
Cross Beam Wind	Night	1.59	.92	-.6	5.1	.79 (.58)	SIGI (ULCOMP)
Speed (SLASER)	Morning	-.92	.40	-2.1	.5	-.53 (.50)	SPEEDT7 (ULCOMP)
	TOTAL	1.53	1.82	-2.1	6.6	.79	ULCOMP
Cross Beam Wind	Day	2.64 (m/s)	2.08	-3.4	6.5	.78	SPEEDT7
Component From Cup	Night	.98	.87	-1.1	3.3	.84	SPEEDT7
Anemometer at	Morning	-1.20	.62	-2.6	.2	-.73	SPEEDT7
Thorn 7 (ULCOMP)	TOTAL	1.48	2.1	-3.4	6.5	.79	SLASER
Thorn 7 Cup	Day	3.50 (m/s)	1.20	.8	7.2	.78	ULCOMP
Anemometer Total	Night	1.39	.65	.4	4.0	.84	ULCOMP
Wind Speed	Morning	1.93	.66	.5	3.9	-.73	ULCOMP
(SPEEDT7)	TOTAL	2.50	1.50	.4	7.2	.73	ULCOMP
Thorn 7	Day	29.3 (°C)	3.3	20.9	34.9	-.43	SPEEDT7
Temperature	Night	23.9	3.5	14.7	30.2	.62	SLASER
(TEMP)	Morning	27.0	2.9	20.1	32.4	-.47	SLASER
	TOTAL	27.4	4.9	14.7	34.9	.53	ULCOMP
Thorn 7	Day	21.1 (°)	11.2	3.2	74.1	-.53	SLASER
Sigma Theta	Night	23.6	17.0	3.8	92.7	-.57	ULCOMP
(SIGI)	Morning	20.7	8.5	8.3	63.2	-.29	SIGI
	TOTAL	21.6	13.6	3.2	92.7	-.32	ULCOMP
Optical Turbulence	Day	.071	.037	0.01	.18	-.34	SPEEDT7
From Std. Dev. of	Night	.082	.038	0.0	.22	.79	SLASER
Laser Intensity	Morning	.082	.020	0.04	.12	-.29	SIGI
(SIGI)	TOTAL	.082	.039	0.0	.22	.25	TEMP

*Day = 1000-2200, Night = 2200-0600, Morning = 0600-1000 PST.

TABLE 2. Statistical summary of σ_θ relationships in the Geysers region using 2224 five minute averaged data with no preselection of drainage wind periods.

Station		UTM Coordinates		Elevation MSL (metres)	Mean σ_θ	Correlation	
#	Name	North	East			Maximum Absolute	Station Name
1.	T.V. Repeater	4292.3	527.0	468.6	23.4	.14	Thorn 7
2.	McKinley 2	4291.5	524.5	740.2	25.1	.19	Thorn 7
3.	Thorn 7	4292.7	524.8	677.4	22.7	.19	McKinley 2
4.	Smith-Brazeale	4294.2	525.0	753.7	25.7	.28	Quonset Hut
5.	PDC-1	4293.6	523.4	908.8	17.5	.17	McKinley 2
6.	Anderson Ridge	4293.3	522.0	1112.8	24.4	.26	Quonset Hut
7.	Quonset Hut	4294.5	525.8	692.9	16.6	.28	Smith-Brazeale

TABLE 3. Summary of results of multiple linear regression applied to the σ_θ station measurements.

Station (Dependent Variable)		Period (5 min. avg.)	Station # Included (F-test 0.05)	% Variance Explained	% Variance Explained Per Station	Mean σ_θ ($^\circ$)
T.V. Repeater	1st	400	3,6	2.7	1.3	21.9
	2nd	400	2,7	4.8	2.4	30.0
	3rd	400	2,3	9.7	4.8	22.5
	4th	400	3,6	4.9	2.4	23.4
	5th	400	3,4	3.0	1.5	21.0
	Day	1st	2	9.6	9.6	29.1
		2nd	2,3,5,7	12.1	3.0	23.2
	Night	1st	3,4,5	6.1	2.0	21.7
		2nd	3	1.8	1.8	22.3
	Morning	384	None	None	None	23.9
	Thorn 7	1st	4,7	8.2	4.1	19.8
		2nd	2,6,7	12.5	4.1	26.3
		3rd	1,4,6,7	16.6	4.1	22.8
		4th	1,2,4,5	27.3	6.8	19.0
		5th	1,4,5,7	7.9	2.0	20.2
	Day	1st	4,6,7	21.4	7.1	23.4
		2nd	1,2,4,5,7	22.5	4.5	21.2
	Night	1st	1,4,7	11.8	3.9	22.3
		2nd	1	1.8	1.8	19.2
	Morning	384	5,6	7.8	3.9	23.6
Anderson Ridge	1st	400	1,2,4,7	16.0	4.0	9.9
	2nd	400	3,5,7	19.5	6.5	28.7
	3rd	400	2,5	18.5	9.2	36.6
	4th	400	2,4,5,7	28.1	7.0	29.9
	5th	400	2,4,7	19.7	8.6	24.1
	Day	1st	2,3,7	27.6	9.2	25.3
		2nd	2,4,5,7	20.1	5.0	27.4
	Night	1st	2,4,5	19.6	6.5	21.6
		2nd	2,3,5	13.2	4.4	15.1
	Morning	384	3,5,7	20.1	6.6	35.1

FIGURE 1. A topographic map of the Geyser's region with the laser path located between the light transmission point (Thorn 7) and receiver location (Aminoil Pipeyard).

FIGURE 2. Topographic relief below the laser path.

SIDEVIEW OF LASER PATH IN GEYSER'S REGION

FIGURE 3. Comparison of time history plots of the optical anemometer and cross-path wind speeds at Thorn 7 for ten day period beginning 18 July 1979 (10 minute average values).

FIGURE 4. Scatter plot of 400 values of the variables shown in Fig. 3 chosen to exclude the thunderstorm period.

Time lagged correlation function
optical anemometer versus Thorn 7 cross wind component

Time lagged correlation function
optical anemometer versus Thorn 7 sigma theta

FIGURE 5. Correlation functions for the two variables shown in Fig. 3 and between the optical anemometer and σ_θ measured at Thorn 7 (10 minute averages).

FIGURE 6. Comparison of normalized spectral densities for the variables in Fig. 3 smoothed with a 5 hour triangular filter.

FIGURE 7. Autospectral densities for wind speed and the cosine of the wind direction from perpendicular to the light path (10 minute averages) smoothed with only a 50 minute triangular filter.

FIGURE 8. Analogous spectral plots to Figs. 6 and 7 for data taken across a windy pass 3 km East of LLL for June and July 1977 (30 minute averages).

FIGURE 9. Comparison of spectra for variables shown in Fig. 6 for data constructed of only 8 hour nighttime values excluding the thunderstorm period.

ASCOT1 DAY NO DRAIN LSA VS X WIND

RAW MODES= 165 TRIANGLE NTS M= 3 DEGF= 1 DF= 4.38E-01 FNYQ= 7.20E+01 ZEROth MODE= 7.965E-02
 NOTE -A TRUE CENTERED SMOOTHING WINDOW OVER M RAW MODES DOES NOT INCLUDE M/2 MODES AT EACH END OF THE SPECTRUM
 THUS THE M/2 END POINTS OF ALL SPECTRAL PARAMETERS ARE OF UNDETERMINED RELIABILITY

FIGURE 10. Autospectral density of 8 hour nighttime values for only the thunderstorm period.

FIGURE 11. Autocorrelation functions based on 10 minute averages of wind speeds shown in Fig. 3.

COMPARISON ENERGY DISIP. LSA VS THORN7 X-COMP 10MIN AV

COMPARISON ENERGY DISIP. LSA VS THORN7 X-COMP 10MIN AV

FIGURE 12. Time truncated average estimates of energy dissipation for point and spacially-averaged wind variation (10 min. average variation over 1

COMPARISON EPS LASER WITH SIGMA THETA THORN7 10MIN AV.

EPS THORN7 WIND SPD VS SIG THETA 10MIN AV.

FIGURE 13. Time history plot of σ_θ measured at Thorn 7 and correlation function with ' estimated from the laser measurement.

RAW MODES= 704 TRIANGLE MYS M= 3 DEGF= 1 DF= 1.02E-01 FNYQ= 7.20E+01 ZEROth MODE= 2.592E+00
 NOTE -A TRUE CENTERED SMOOTHING WINDOW OVER M RAW MODES DOES NOT INCLUDE M/2 MODES AT EACH END OF THE SPECTRUM
 THUS THE M/2 END POINTS OF ALL SPECTRAL PARAMETERS ARE OF UNDETERMINED RELIABILITY

FIGURE 14. The autospectral densities for the values of σ_θ at Thorn 7 shown in Fig. 13.

**Observations of Nighttime Winds Using
Pilot Balloons in Anderson Creek Valley,
Geysers, California**

by

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and
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Abstract

Nighttime winds in Anderson Creek Valley located in the Geysers area of northern California are examined using pilot balloons as air parcel tracers. Observations show a persistent night-to-night pattern of winds above Anderson Creek Canyon and in the valley basin. Before sunset, strong down-slope winds exist; these are believed due to terrain interactions with the late afternoon sea-breeze. The drainage wind within the creek canyon has an average speed of about 3 m/s, grows in thickness at a rate of about 80 m per kilometer of downwind distance and is insensitive to above-valley wind speeds. Balloons in the drainage wind quickly spread throughout the creek canyon. The transition from valley winds to the free, above-valley winds occurs at about 500 m above the valley floor for weak above-valley winds, and about 300 m for strong above-valley winds.

1. Introduction

Nighttime winds in a valley are complicated mixtures of drainage winds along the valley slopes and sidewall canyons and terrain induced flows resulting from the interaction of above-valley winds with the valley ridges. In a complex valley, the wind system is spatially variable and hence fixed point observations fail to give a complete picture of the flow structure. Because of this, many studies of valley winds and winds over complex terrain have been made by observing the motions of zero-lift or slow-rising balloons (see e.g. Gifford 1953, Buettner and Thyer 1965, and Wooldridge and Orgill 1978).

This report presents the results of a series of nighttime wind observations made using slow-rising balloons in the Anderson Creek Valley located in the Geysers area of northern California. These observations were part of the July, 1979 field study conducted in this valley by the Department of Energy's Atmospheric Studies in Complex Terrain (ASCOT) program (see Dickerson and Gudiksen, 1980). The valley, pictured in Figure 1, lies within the Mayacmas Mountains and is bounded by Cobb Mountain on the northwest and by Boggs Mountain on the northeast. The valley opens to the southeast. During the night, valley winds are from the northwest and generally follow Anderson and Putah Creek.

Valley sides along Anderson Creek have slopes of about 10° to 15° over distances of about 3 km. Anderson Creek lies at the bottom of a shallow canyon which is initially narrow and steep-walled but broadens and flattens after joining Gunning Creek. Along Anderson Creek, the valley sides are covered with bushes (~1.5 m), tall grass (~0.5 m), isolated trees (~10 m) and rock outcroppings. The valley floor, especially around the community of Anderson Springs is thickly forested.

2. The Data

During four 6 P.M. - 6 A.M. observation periods, slightly buoyant 30 gm pilot balloons were released near Anderson Creek about halfway between the creek headwaters and the spot where the creek slope levels off. The release point is about 230 m above the valley floor and about 500 m below the east-facing ridge of the valley. A double theodolite system with a baseline of 490.7 m was used for balloon tracking. The theodolite baseline extended across the creek canyon in a north-south line about 100 m east of the release point. Both theodolites were located on high (~30 m) rock outcroppings so that balloons were visible even when deep in the creek canyon. Elevation difference between theodolites were less than 1 m. Theodolite readings were made at 30 sec intervals. Water-activated battery-and-lamp systems were fixed to the balloons for nighttime tracking. About 25

balloons were released per night, and average tracking time per balloon was about 20 minutes. Detailed balloon trajectory, temperature profile and wind data are given in Nappo et al. (1980).

At the beginning of this study, it was observed that zero-lift balloons would sink to the ground or be caught in trees within a minute or so after release. If, however, a small lift, corresponding to a rise rate between roughly 0.25 and 0.50 m/s, was given to the balloons, they would drift above the trees yet remain in the drainage wind layer. To observe winds above the drainage layer, slow rising balloons with rise rates of about 1 m/s were used. Above-valley winds were observed by releasing balloons with rise rate of about 3 m/s. Except for the rapidly rising balloons, all balloons stood for about 10 to 20 minutes in an open van in order to reach thermal equilibrium with the ambient air.

3. Results

3.1 Terrain effects

Most of the balloons released before sunset traveled in an easterly direction at relatively high speed (5 - 7 m/s) with little change in altitude. After a few minutes of travel, some were observed to rapidly slow to a speed of a few tenths of a meter per second while rising at a rate of about 1 m/s. In many instances, rising balloons drifted in an up-valley direction, e.g. back toward the release point. An example of these motions is given in Figures

2 and 3 which show the horizontal and vertical trajectories of a typical late afternoon balloon release. These motions suggest a terrain interaction with the late afternoon sea breeze resulting in lee waves. Studies of the sea breeze in Central California by Fosberg and Schroeder (1966), and Schroeder et al. (1967) show that the sea breeze often extends into the Geysers area. These wind patterns were not evident for very long after sunset.

3.2 The drainage wind

The drainage winds above Anderson Creek became fully developed within a few hours after sunset. Table I lists the times of the onset of drainage, the average wind speed within the drainage layer, and the winds measured at about 1000 m above the release point i.e., the above-valley winds. The average speed of the drainage wind is relatively unaffected by the above-valley winds. The time for the development of the drainage on Anderson Creek varies from 1900 to 2200 (PST).

The drainage wind patterns generally remained the same from night to night. Path-averaged drainage winds ranged from 1 to 5 m/s with an ensemble average speed of about 3 m/s. On reaching the valley floor, the drainage wind usually diminishes to speeds less than 1 m/s. Horizontal and vertical trajectories for a typical balloon are seen in Figures 4 and 5 respectively. The ensemble of vertical trajectories of

all drainage wind balloons is given in Figure 6. The terrain plotted in this figure is taken along the mean horizontal trajectory line. The drainage wind layer is about 50 m deep at 300 m and about 100 m deep at 900 m distance, indicated a growth rate of about 80 m/km. In an analysis of two-dimensional slope flows based on the model of Ellison and Turner (1959), Briggs (1980) shows that for the case of constant cooling rate at the surface and slope inclinations less than 20° , the drainage wind layer thickness, h , is linearly related to fall distance, x , by the relation $h = 0.037 x \beta^{2/3}$ where β is the slope angle. The average terrain slope above Anderson Creek is about 10° which results in a predicted growth rate, h/x of about 12 m/km. The drainage layer is seen to grow at a rate some seven times greater than that predicted for a two-dimensional slope.

The crosswind dispersion of balloons within the drainage layer at 250, 500, 1000, and 1500 m distance is shown in Figure 7. The view is toward the release point. Crosswind distances are measured from the centroid of the balloon positions. In general the balloons seem to disperse across the width of the canyon at all distances downwind.

3.3 The transition height

During the experiment, it was commonly observed that slow rising balloons would experience a sharp increase in horizontal

and vertical speeds a few hundred meters above the valley floor. This discontinuity in wind speed is believed to mark the transition from local valley winds to above-valley winds. This is demonstrated in Figure 8, which shows the vertical trajectories of two balloons released within an hour of each other. The balloon released at 2337 (PST) was balanced so as to drift with the drainage wind. The balloon released at 0020 (PST) was balanced to rise slowly. At a height of about 100 m above the release point, both balloons indicate an increase in horizontal wind speed. On this particular night, the above valley winds were strong as shown by the wind profiles in Figure 9. These wind profiles were measured with free-lift pilot balloons. Analyses of similar trajectories show that for light above-valley winds (2 - 3 m/s) the transition height is about at ridge height, i.e. about 500 m above the valley floor, but during strong above-valley winds (5 - 7 m/s) the transition height is about 300 m.

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Much of the data reduction and computer analysis were done

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TABLE I

DATE (July '79)	DRAINAGE ONSET TIME (PST)	AVERAGE DRAINAGE SPEED (m/s)	ABOVE VALLEY WINDS	
			SPEED (m/s)	DIRECTION (deg.)
18	1900	3	<2	345°
22	2000	3	6	250°
24	2200	3	2	190°
26	2100	2	1	300°

FIGURE 1. Anderson Creek Valley region. Contours are in m (MSL).

FIGURE 2. Horizontal trajectory of balloon released at 1817 PST on 26 July 1979. Balloon positions are marked at 30 second intervals.

FIGURE 3. Vertical trajectory of balloon released at 1817 PST on 26 July 1979. Balloon positions are marked at 30 second intervals.

FIGURE 4. Horizontal trajectory of balloon released at 0459 PST on 19 July 1979. Balloon positions are marked at 30 second intervals.

FIGURE 5. Vertical trajectory of balloon released at 0459 PST on 19 July 1979. Balloon positions are marked at 30 second intervals.

FIGURE 6. Vertical trajectories of all balloons in the drainage wind.

ORNL-DWG 80-8335

FIGURE 7. Crosswind positions of balloons in the drainage wind.

ORNL-DWG 80-11844

FIGURE 8. Vertical trajectories of balloons released at 2337 PST on 22 July and at 0020 PST on 23 July 1979. Balloon position are marked at 30 second intervals.

FIGURE 9. Above-valley winds during morning of 23 July 1979.

NET RADIATION, VERTICAL PROFILES OF WIND AND TEMPERATURE,
AND CROSS-VALLEY FP TRACER SAMPLING IN THE
ANDERSON CREEK VALLEY AREA - JULY 1979

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ABSTRACT

The Department of Atmospheric Sciences of PNL assisted the ASCOT program during its July 1979 field effort by collecting data from a 30 m tower, tethersonde, monostatic acoustic sounder, FP sampling stations, surface wind sensors and two net radiometers. Preliminary results from the net radiometers, PNL (and WPL) tethersonde, and the FP tracer sampling will be discussed in this report.

INTRODUCTION

The purpose of this report is to describe Pacific Northwest Laboratory's (PNL) contribution to the ASCOT field project between July 15 to 28, 1979. PNL's field effort consisted of the following tasks:

- 1) collecting wind, turbulence and temperature data at four levels from a 30 m tower located at the entrance of Anderson Creek Valley (Diamond D Ranch); soil temperatures and net radiation measurements were also obtained;
- 2) collecting temperature, humidity, pressure and wind data up to about 1000 m with a mobile tethersonde;
- 3) operating a monostatic acoustic sounder for obtaining data on the height and structure of temperature inversion and thermals;

- 4) establishing and operating a network of five sampling stations for sampling fluorescent particles (FP) tracer released in the upper part of Anderson Creek;
- 5) collecting supplemental net radiation and surface wind data.

The general topography of Anderson Creek and the location of the various field operations is shown in Figure 1. Many of the field operations, with the exception of the tower, were conducted at PDC-1 well site at approximately 840 m above sea level. On the evening of July 16 a tether sonde was operated at the Diamond D Ranch location in conjunction with the tower and the dual doppler acoustic sounder of the Wave Propagation Laboratory (WPL).

The main objective of the field effort was to define wind velocity and temperature structure, drainage layer depth, turbulence, and transport along Anderson Creek drainage basin. Therefore, much of the operation was conducted during the nighttime hours. Figure 2 indicates the chronological time periods in which data were obtained from the various sensors. Thunderstorms during the evening of July 20 and gusty winds during the evening of July 27 limited data collected on those dates. Data collection during the daytime hours were limited because manpower was not available for a 24-hr operation.

This report discusses the preliminary results from three data sets:

- 1) the net radiometer data at PDC-1 and Diamond D,
- 2) the PNL and WPL tether sonde data and
- 3) the cross-valley FP sampling data.

NET RADIATION MEASUREMENTS AT GEYSERS GEOTHERMAL SITE

The radiation balance at the earth's surface is of concern in the development of atmospheric dispersion models. In particular, the rate of surface cooling is an important factor in valley drainage flow. Pacific Northwest Laboratory (PNL) deployed net radiometers^(a) at two valley locations during the ASCOT July 1979 field program.

(a) Temperature-compensated Miniature Net Radiometer manufactured by Micromet Instruments, Bothell, WA.

Although the two sensors used were identical, the methods of processing and recording the sensor outputs were not. The output of the Diamond D radiometer (at a down-valley location) was amplified, scanned once per second, digitized and recorded on a 9-track magnetic tape. The taped digital data were processed by computer with the output being 15-min mean values of net radiation. The output of the radiometer at PDC-1 (at an up-valley location) was amplified, and the resultant analog signal was recorded on a strip chart. The strip chart was reduced by manually averaging over 10-min increments.

Measurements were not continuous at either site but were centered around selected, intensive, measurement periods. Figure 3 displays the times and magnitude of net radiation measurements for the 13 days of field measurements. The duration of an individual Diamond D experiment was limited by the approximate 19 hr of data that could be recorded on one magnetic tape with the selected 1-sec scan rate. The PDC-1 data were restricted to periods when an electrical power generator was in operation at the site.

Although this note is not intended as indepth examination of the radiometer data, some preliminary comments and observations based on the analog traces of Figure 3 are in order.

The period of net positive radiation is about 12 hr/day at each site, with positive values starting at about 0600 PST and ending shortly after 1800 PST. The daytime net radiation at the Diamond D site is greater than that at the PDC-1 site on all occasions where comparison is possible. This observation suggests a larger albedo at the PDC-1 site.

At night, the net outward flux of radiation from PDC-1 was slightly greater than at Diamond D for the first four comparisons, i.e., through July 25; but the last two comparisons reverse this trend. Pertinent to this reversal is the fact that the PDC-1 radiometer was moved from over a gravel surface to over a grassy surface on the afternoon of July 26. The change in character of nighttime net radiation following this instrument relocation could possibly result from a thermally insulating effect of the vegetation and the porous, grassy surface compared to the previous compact, bare, gravel surface. Since the daytime outward fluxes on July 27 and 28 remain greater at Diamond D than at PDC-1

it appears that the albedo of the PDC-1 site remained greater than that at Diamond D. The surface beneath the Diamond D net radiometer also had grass and weeds--a vegetative cover in the fenced area of the plowed field where the Diamond D instrumentation was exposed.

The fluctuation of the radiation balance near sunset at Diamond D on July 16, 18 and 20 likely resulted from trees to the northwest of the radiometer. The fluctuation diminished on July 20 and was not observed on subsequent dates. This change likely reflects the changing azimuth of the setting sun.

Most nights were essentially cloud free. A dramatic exception was on the evening and night of July 20. Skies were overcast with thunderstorms in the area. The cloud cover resulted in reduced outgoing radiation that night.

Although only analog traces of the net radiation data are presented here, data from Diamond D and from PDC-1 are available in digital form.

TIME VARIATION OF WIND AND TEMPERATURE FROM TETHERSONDE DATA

Two mobile tethersonde units were utilized in the Anderson Creek area near Middletown, California during the July, 1979 field effort for the ASCOT program. One unit was from Pacific Northwest Laboratory (PNL) and the other was from the Wave Propagation Laboratory (WPL) in Boulder, Colorado. On the evening of July 16, a cooperative effort between the two groups obtained tethersonde data at the Diamond D Ranch site for 18 hr. After that period, the PNL tethersonde was located at the PDC-1 well site in the northwestern part of the Anderson Creek drainage (840 m) for the duration of the project. WPL's tethersonde was located at the Thorne 7 well site (600 m) around 1.1 km southeast of the PDC-1 well site (Figure 1).

Data were obtained from both tethersonde sites for six nighttime and early morning periods. Vertical profiles of wind, temperature and humidity were taken about every hour except when strong winds, equipment malfunction or weather prevented it. Ground stations received the data remotely, which was printed on a paper tape.

The PNL and WPL tethersonde data volume was edited and computer plotted as well as being placed on punch cards and a magnetic data tape. The edited, nonaveraged data were processed by a computer plotting routine, as shown in the example of Figure 4, for incorporation into the ASCOT Data Summary (1980) and for preliminary assessment and analysis.

A number of operational problems were encountered in using the tethersonde system which are noted below yet the data volume as a whole made an important contribution to the field project by showing the time and height variation of the winds, temperature and humidity structure over the Anderson Creek area. A few months after the ASCOT field program it was discovered that the tethersonde pressure sensor was sensitive to physical handling which ultimately could affect the altitude readings. Although the editing program attempted to eliminate obvious bad data some questionable altitude data still was accepted and plotted, especially on descent profiles. After this problem was found corrective action and new pressure sensors has helped alleviate these height errors.

Another problem with the tethersonde system is with the wind direction sensor. If winds are varying around due north (360°) the wind direction sensor can give directions greater than 360° . This problem is inherent with the present system and was not corrected by the editing program. This was not a particular problem in this preliminary analysis but must be recognized when the data is analyzed in detail. Three other problems should be briefly mentioned: 1) WPL tethersonde humidity sensor did not give correct humidity values all the time and the obvious bad data were not plotted by the editing and plotting routine. 2) During relatively strong winds ($> 8 \text{ m s}^{-1}$) the tethersondes would not plane up into the wind because of the wind drag. Thus, the tethersondes could not reach very high altitudes and were not directly over the launch site. Trees surrounding the launch sites also caused problems for the tetherlines during periods of strong winds. 3) The temperature sensors have a 10 sec response time resulting in a slight hysteresis effect in the temperature data.

Generalized temperature and wind speed vertical profiles for afternoon and nighttime hours as deduced from the tethersonde data volume is depicted

in Figure 5. Figures 6 to 14 present a preliminary presentation of the time variation of temperature structure and winds for the experimental nights of July 16-17 (Diamond D), 18-19, 22-23, 24-25, and 26-27. Two experimental nights July 20-21 and 27-28 were not included because of weather conditions (thunderstorms on the 20th and strong winds on the 27th) and insufficient data. Wind data are only shown by west (ripples) and east (no ripples) components. Temperature structure were divided into six different stability classes as depicted in Figure 5.

Generally, during the experimental periods, late morning easterly upslope winds were replaced by westerly downslope winds early in the afternoon (~ 12 to 14 p.m.). These afternoon westerly winds were apparently related to the combined monsoonal-sea breeze and west-side valley-slope wind that is apparently common in this area in the summertime (Schroeder 1961). The afternoon westerly winds exhibited peaked vertical wind speed profiles with low-level (50 to 400 m abg) wind maxima (LLSM) varying from 3 to 11 m s^{-1} . These wind speed maximas usually decrease in magnitude by midnight for most experimental nights. (a) The atmospheric stability in the lower 500 to 600 m was usually neutral or unstable during these afternoon westerly downslope winds (see Figure 5).

Time variation of vertical profiles of wind and temperature was strongly influenced by the diurnal net radiation cycle (Figure 3). With sunset and the onset of negative net radiation the lower atmosphere in the Anderson Creek area quickly cooled forming as many as four distinct stable layers (Figure 5). Generally, these were: 1) a shallow (50 to 100 m) very stable layer near the ground ($+0.1$ to 0.2°C/m), 2) a primary stable layer ($\sim +0.02$ to $+0.03^\circ\text{C/m}$), 3) a isothermal layer (bottom and top of temperature inversion) and 4) a slightly stable layer. Quite frequently a near neutral (adiabatic) layer was found above the temperature inversion or the slightly stable layer.

On first examination of tethersonde data the nocturnal drainage winds were apparently shallow and poorly defined. Possibly one reason for this

(a) Figures 9 and 10, July 22-23, 1979 was an exception.

was the superposition of the decaying afternoon westerly wind (Figures 6 and 8).

Best evenings for apparent drainage wind phenomena appeared to be July 16-17, 18-19, 24-25 and 26-27. Depth of the drainage layer on the slope varied with time and probably did not exceed 100 m very often. Drainage wind speed maxima were 2 to 4 m s⁻¹. Periods of easterly winds above the low-level nocturnal westerly downslope winds were frequent on most nights. The night and early morning of July 26-27 was exceptional in this regard (Figures 13 and 14). Some of these easterly winds may be anti-mountain winds as defined by Buettner and Thyer (1966) and some may be related to upper-level gradient flow but this is conjecture at this point in the analysis.

On those experimental mornings that observation were taken 3 to 4 hours beyond sunrise, the nocturnal stable conditions and the westerly drainage layer were observed to quickly (1 to 2 hrs) dissipate with the onset of positive net radiation. Daytime neutral and unstable stability conditions were established on the slopes as well as easterly upslope winds (Figure 8, 9 and 10).

FLUORESCENT PARTICULATE TRACER EXPERIMENTS AT GEYSERS GEOTHERMAL SITE

Among the field experiments performed during this period were atmospheric tracer releases. Personnel from Los Alamos Scientific Laboratory (LASL) released fluorescent particulate (FP) tracer during 4 nights when drainage flow was anticipated. Tracer release was from a point near the head of Anderson Creek at an elevation of about 1140 m (3730 ft) msl. Tracer releases lasted from 21 to 25 min, whereas down-valley sampling persisted for a minimum of 2 hr at each of the five LASL sampling locations. The LASL sampling was performed at near-stream-bed locations at distances 2.7 to 7.6 km from the release point. At three of the five LASL sampling sites, samplers were operated in sequential fashion in order to obtain a measure of the history of tracer concentration. Integrating samplers were operated at all five sites. The samplers used by LASL were Rotorods.

Pacific Northwest Laboratory (PNL) complemented the LASL tracer sampling by operating membrane filter samplers at five sampling locations. Whereas the LASL sampling was designed to define down-valley profiles of FP concentration, the PNL sampling was designed to examine the cross-valley profile. The

PNL samplers were located near the intersection of Anderson Creek and Socrates Mine Road, a distance about 2.7 km from the release point (Figure 15). An integrated sample as well as sequential samples were collected at the creek-road intersection (at the same location as an LASL sampling station), but unattended integrating samplers only were operated at four stations north and south of the creek bed. Greater detail on sampler location is shown in Figure 16. Figure 17 gives a cross-valley profile of elevation along a roughly north-south line near the PNL samplers. Although it would have been preferable to space samplers evenly with respect to elevation or distance, dense vegetation and steep terrain made such a sampling layout impractical. Sample locations (circles) on Figure 16 were plotted on the basis of elevation (not north-south distance from the Anderson Creek center position).

Down-valley drainage was suggested during three of the four tracer releases near the head of Anderson Creek. Results of these releases are shown on Figures 17, 18 and 19. The magnitude of dosage (time-integrated concentration) isoplethted on these figures has been normalized to be compatible with LASL Rotorod counts. Although the displacement of the maximum dosage to the south wall of the valley is not explained, the data are generally suggestive of a drainage down Anderson Creek. Dosages at the valley floor are 75% or more of the observed maximum. Winds showed convergence toward Anderson Creek from both valley walls during tracer passage at this location.

In order to be able to derive at least a "ball-park" estimate of the lateral spread (σ_y) of the tracer plume, the concentration profiles for tests 1, 3 and 4 were extrapolated to the south. Although an attempt was made to give an appearance of symmetry in the extrapolation, the procedure was very subjective. The σ_y values derived for Tests 1, 3 and 4 were, respectively, 71, 122 and 144 m--values that are commensurate with those found in flat terrain at night for a similar downwind distance.

The release of 22 Jul 79 showed a different ground-level dosage pattern (Figure 20). The dosage at the valley floor (650 m msl) was less than a third of the maximum observed near the top of the north valley wall (728 m msl). Background tracer dosage was observed just 30 m up the south wall of the valley. It is postulated that the trajectory of tracer was not down the Anderson Creek

Valley, but instead tracer moved in an ENE or E direction from the tracer source toward PDC-1 and Ford Flat. The appearance of tracer at the PNL sample transect suggests a drainage from the north into the valley. Figure 21 presents an areal plot of concentration based on both PNL filter and LASL Rotorod data. Observed wind directions are also shown. The winds in the area between Anderson Creek and Gunning Creek are in agreement with this postulated flow.

Figure 22 presents the dosage pattern of 18 Jul 79 - one of the patterns more compatible with drainage flow down the Anderson Creek canyon. LASL Rotorod data suggest the plume centerline was considerably closer to the samplers in Anderson Creek valley (sampler 12 and 13) than to samplers at comparable distances from the source in Putah Creek valley (17 and 18). By contrast, returning to Figure 21 we see that the postulated trajectory following the 22 Jul 79 release at the head of Anderson Canyon led to concentrations high on Putah Creek valley than in Anderson Creek valley.

It is perhaps interesting to note that the peak dosage measured at the PNL sampler farthest up the Anderson Creek north wall (i.e., closest to Ford Flat) was nearly double the dosage measured following any of the tracer releases. Tracer samplers at PDC-1 and along Ford Flat Road northeast from PG&E met station 19 would have been very instructive in the analysis of the release of 22 Jul 79. It is recommended that since these are areas of relatively easy access, sampling for the Anderson Creek tracer(s) should be carried out here during the 1980 experimental period.

Figure 23 presents the results of the sequential sample at the Anderson Creek-Socrates Mine Road crossing. Sequential sampling increments were 15 min during Test 1 and 10 min during the other three tests. A comparison of the duration of tracer releases (as depicted at the left for each test) with the duration of tracer passage at this 2.7-km down-valley point reveals well-above background concentrations for a period of at least twice the dispersal duration. The dotted lines, derived from the integrating samplers, indicate the concentration expected from a continuous plume. Note that only during Test 2 did the observed short-period concentration reach that expected continuous plume value. The "x" symbols on Figure 23 represent

the concentrations derived from LASL Rotored samplers--samplers operated over the same time increments as the filter samplers.

On the basis of the 1979 experiments, it can be concluded that if crosswind tracer sampling is included in subsequent ASCOT field experiments, sampling should be extended much farther up the side of the valley.

CONCLUSIONS

This report has been a preliminary assessment of the field data that PNL and other groups (WPL, LASL) collected during the ASCOT field experiment in July 1979. Any conclusions made now are very tentative since detail analysis of all the data volume should present a more coherent picture of the meteorological and dispersion phenomena of this period.

This part of PNL's contribution was to provide analysis of net radiometer, tethersonde, and FP tracer data. In general, the net radiation data, although not covering the full diurnal cycle, revealed a fairly typical radiation cycle for a summer period without clouds. Tethersonde soundings, on the slopes and in the valley, showed the winds and temperatures over the Anderson Creek area were strongly influenced by the net radiation cycle. Although nocturnal drainage winds were apparently observed on the slopes and in the valley they were shallow and, at times, apparently influenced by the synoptic monsoonal-sea breeze airflow pattern that is present in this area in the summertime. FP tracer experiments also revealed the presence of nocturnal drainage winds although the experimental test of July 22-23 appeared to be a little anomalous possibly due to synoptic wind influences.

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FIGURE 1. GENERAL TOPOGRAPHY OF ANDERSON CREEK AND LOCATION OF ASCOT FIELD OPERATIONS DURING JULY 15-28, 1979

FIGURE 2. PNL'S DATA COLLECTION TIME PERIODS DURING ASCOT FIELD OPERATIONS: JULY 1979

- 1) PDC-1: Strip chart pen problem.
- 2) Diamond D: Sensor moved to new location to avoid mid-afternoon tower shadow; moved before start of data collection on July 20, 1979.
- 3) PCD-1: Strip chart jammed from about 1645 to 1940.
- 4) PDC-1, 1230: Change from nitrogen purge gas to helium.
- 5) PDC-1, 0531: "Sun just up over mountains to east."
- 6) PDC-1, 0400: "Dawn light" visible.
- 7) PDC-1, 0614: "Cloudy. Altocumulus scattered to the east."
- 8) PDC-1, 1554 to 1607: "Moved radiometer to different location." The radiometer was over a gravel surface and was moved to "grassy surface."
- 9) PDC-1, 0930: "Weather clear."

FIGURE 3. NET RADIATION VERSUS TIME FOR DIAMOND "D" AND PDC-1 LOCATIONS DURING ASCOT FIELD EXPERIMENTS: JULY 1979

TETHERSONDE (PNL) AT PDC-1

DATE: 072379 0333 TO 0418 PST (TOP 0403)

TETHERSONDE (WPL) AT THORNE 7

DATE: 072379 0333 TO 0417 PST (TOP 0358)

FIGURE 4. AN EXAMPLE OF THE PNL AND WPL TETHERSONDE DATA FOR JULY 23, 1979 AS REPRODUCED BY A COMPUTER EDITING AND PLOTTING ROUTINE

FIGURE 5. GENERALIZED TEMPERATURE AND WIND SPEED VERTICAL PROFILES FOR AFTERNOON AND NIGHTTIME HOURS AS DEDUCED FROM THE TETHERSONDE DATA VOLUME

FIGURE 6. TIME VARIATION OF TEMPERATURE STRUCTURE AND WIND PROFILES. DIAMOND D, JULY 16-17, 1979. 1701 TO 1003 PST. SEE FIGURE 5 TEXT FOR KEY.

FIGURE 7. SAME AS FIGURE 6, BUT FOR PDC NUMBER 1, JULY 18-19, 1979. 1713 TO 0050 PST

FIGURE 8. SAME AS FIGURE 6, BUT FOR THORNE 7, JULY 18-19, 1979. 2013 TO 0845 PST

FIGURE 9. SAME AS FIGURE 6, BUT FOR PDC NUMBER 1, JULY 22-23, 1979. 1707 TO 0925 PST

FIGURE 11. SAME AS FIGURE 6, BUT FOR PDC NUMBER 1, JULY 24-25, 1979. 1659 TO 0653 PST

FIGURE 12. SAME AS FIGURE 6, BUT FOR THORNE 7, JULY 24-25, 1979. 1908 TO 0601 PST

FIGURE 13. SAME AS FIGURE 6, BUT FOR PDC NUMBER 1, JULY 26-27, 1979. 1808 TO 0634 PST

FIGURE 14. SAME AS FIGURE 6, BUT FOR THORNE 7, JULY 26-27, 1979. 1855 TO 0634 PST

FIGURE 15. LOCATION OF THE PNL CROSS-VALLEY FP SAMPLING NETWORK, RELEASE POINT OF TRACER, AND LASL FP SAMPLING LOCATIONS

FIGURE 16. CROSS-VALLEY FP SAMPLING NETWORK SHOWING LOCATIONS OF INDIVIDUAL SAMPLERS

FIGURE 17. CROSS-VALLEY PROFILE OF ELEVATION ALONG FP SAMPLING NETWORK. FP SAMPLING RESULTS OF JULY 18, 1979. RELEASE 2300 TO 2325 PST. (TEST 1) COUNTS COMPATIBLE WITH LASL ROTOROD COUNTS

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FIGURE 18. FP SAMPLING RESULTS OF JULY 24, 1979. RELEASE 2225 TO 2246 PST. (TEST 3)

FIGURE 19. FP SAMPLING RESULTS OF JULY 22, 1979. RELEASE 2215 TO 2238 PST. (TEST 4)

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FIGURE 20. FP SAMPLING RESULTS OF JULY 22, 1979. RELEASE 2205 TO 2238 PST. (TEST 2)

FIGURE 21. DOSAGE PATTERN FROM FP RELEASE OF JULY 22, 1979. ARROWS ARE NEAR-SURFACE WIND DIRECTIONS

FIGURE 22. DOSAGE PATTERN FROM FP RELEASE OF JULY 18, 1979. ARROWS ARE NEAR-SURFACE WIND DIRECTIONS

FIGURE 23. FP TRACER CONCENTRATION VERSUS TIME AT ANDERSON CREEK - SOCRATES MINE ROAD CROSSING

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ANALYSIS OF FLUORESCENT PARTICLE TRACER DATA

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ABSTRACT

Four fluorescent particle tracer experiments were conducted during the July 1979 ASCOT experiment in the Anderson Creek Valley of northern California. The purpose of the experiment was to examine the transport and elongation of a plume traveling in the Anderson Creek nocturnal drainage flow and investigate the interaction of the Anderson Creek and Putah Creek flow fields. Sequential samples of tracer material at three downwind locations in Anderson Creek gave effective transport velocities of 1 to 2 m/s and showed an approximately linear relationship between plume elongation and travel distance. Integrated samples taken in both the Anderson Creek and Putah Creek air sheds indicated considerable interaction between the two flow fields.

I. INTRODUCTION

In the July 1979 ASCOT intensive field study in Anderson Creek Valley, California fluorescent particle (FP) tracer experiments were conducted during drainage flow conditions on four of the experimental nights. The objectives of these were:

- (1) To investigate the transport time and elongation of a plume of material carried down valley by the nocturnal drainage wind in both the Anderson Creek and Putah Creek air sheds.

(2) To investigate any possible interactions between the two air sheds.

This paper reports on the analysis of the tracer data collected. The unknown deposition properties of the FP material negates any quantitative calculation of absolute concentrations; however, for meeting the objectives of this study the FP tracer proved to be a good and simple one to use.

II. EXPERIMENTAL DESIGN

In each experiment tracer was released in the upper portion of each air shed and both sequential and integrated samples collected at downwind locations. The release sites and sampling locations are shown in Fig. 1. Detailed locations of each are given in the Appendix. In the first experiment (FP-1) the position of R1 was located slightly further below the ridge line than in the rest of the experiments in order to insure release in the drainage flow.

In the fourth experiment (FP-4) the Putah Creek release was made further north at location R3 near Whispering Pines (See Appendix and Fig. 2d) to examine the transport from that location. Yellow FP was released in Anderson Creek (R1) and green FP was released in Putah Creek (R2, R3) during each experiment. The FP material was obtained from Ted Brown Associates (Los Altos Hills, CA) and the characteristics of each are given in Table I. A total of 500 g was released over a period of approximately 20 min at each release site. Table II gives the actual start time and duration for each release. Releases were made with a standard FP aerosol generator (Metronics Associates, Palo Alto, CA; model 10) operated off of portable 110 VAC generators. All samples were collected on type H Rotorods using a Model 60A Rotorod Sampler (both Metronics Associates). The sampling rate with this configuration is 52 liters per minute with a collection efficiency of 65%.

The integrated sampler ran for sufficient time to allow for complete plume passage. The sequential sampling was begun as close to the time of release as possible and continued for 2 to 4 h depending on location with

samples taken for 10 or 15 min intervals. Assay of the collection rods was done by visual counting of particles on random representative areas of each while illuminated with an ultra-violet light source. The total number of particles collected is then calculated based on the ratio of the area counted to the total collection area of the rod. Normally the assay was made by two individuals and the results averaged to produce the reported value.

III. RESULTS

A. Integrated Samples

The results of the integrated sampling network for the four experiments are shown in Figs. 2a-d. The total particles collected at each sampler are shown as a logarithmic bar graph near each sampling location. The shading of the bars are matched to the shading at the release point to identify the two tracers. The integrated values displayed at sampler 10 were derived from the sequential samples taken at that location.

There are several observations that can be made from the integrated data:

(1) The total particles collected at site 1 was remarkably constant in all four experiments with those collected during FP-1 being somewhat higher than the other three experiments.

(2) Tracer material released from R1 at the top of Anderson Creek was collected at every sampler in all four experiments.

(3) Tracer material released from R2 at the top of Putah Creek was collected in all samplers except at site 1 (it wouldn't be expected there) in all experiments except FP-2.

(4) In experiment FP-2 large amounts of R1 tracer were found in the Putah Creek samplers with comparatively low to no amounts of R2 tracer in all samplers.

(5) In all experiments site 8 collected more R2 tracer than the closer site 7, except in experiment FP-2 when site 8 collected no R2 tracer.

Based on these limited observations it would appear that:

(1) Drainage flow characteristics in the region from release point R1 to sampling site 1 were about the same each night due to the better defined Anderson Creek air shed in the region.

(2) In the lower area of Anderson Creek below sampling site 1 the drainage flow becomes more complex due to the interaction of other air sheds as evidenced by the different distributions of tracer in the integrated samplers in that region on different nights.

(3) The drainage flow characteristics of Putah Creek were not well defined by the R2 releases. This may be due to either a weaker drainage system or that the tracer was not released completely in the drainage flow regime.

(4) In three out of the four tracer experiments there was considerable transport interaction between the Anderson Creek and Putah Creek flow regimes.

(5) In experiment FP-2 there appears to have been a considerable transport of all tracer in an easterly direction as evidenced by high R1 tracer in Putah Creek and very low R2 tracer in all samplers.

It must be noted that the above are based upon the results of only four tracer experiments and, therefore, may not be representative. However, they do indicate quite well the complexity of interacting flows in the basin.

B. SEQUENTIAL SAMPLES

Sequential samples were collected at sampling sites 1, 9, and 10 (Fig. 1). These locations were selected primarily to look at plume transport from R1 down the Anderson Creek air shed. Although the sequential samplers at location 9 and 10 did collect some R2 tracer the counts were in general too low to interpret plume transport characteristics in Putah Creek. Therefore, the ensuing analysis is concerned only with the sequential sampling results for R1 released into Anderson Creek.

Figures 3a-d show the total particles collected in each sequential sampling interval as a function of time after start of release for each of the four experiments. Each curve is labeled with the distance of the sampler from the release point R1 and the duration of the release is shown by a horizontal bar in the upper left-hand corner of each figure. In the first experiment (July 18-19) sampling was terminated prematurely due to the unexpected elongation of the plume at sites 9 and 10.

The results of the sequential sampling for the first three experiments are fairly similar. They show a rather narrow plume (in time) at site 1 with considerable elongation (in time) at sites 9 and 10. The last experiment (July 26-27) shows a similar plume at site 1, but poorly defined ones at sites 9 and 10 indicating much different transport conditions in the lower part of Anderson Creek on that night than during the previous three experimental nights.

Plume arrival time at each sampler location was determined from Figs. 3a-d by determining the time at which the number of particles reached 10% of the peak value at that sampler. The 10% of peak value method was chosen as an objective method to treat each set of data. From the arrival time and the distance from the release point an effective transport velocity V_{eff} was computed in each case. In addition a plume elongation factor was determined by dividing the width (in time) of the plume at the 10% of peak point by the duration of the release. These effective transport velocities

and elongation factors are given for each experiment at each sampler in Table III. Also listed is the elongation factor per kilometer of travel distance.

An inspection of this data reveals that:

(1) The effective transport velocity from R1 to sites 1 and 10 was fairly constant with values ranging from 1.0 to 1.4 m/s excluding the value of 0.6 at site 10 in the fourth experiment due to highly questionable data.

(2) In the first three experiments an effective transport velocity of 2.0 m/s was determined for transport from R1 to site 9 located between site 1 and 10. This would appear to be an anomaly. This result implies that the plume was transported faster between sites 1 and 9 and then slowed down again between sites 9 and 10. Another possible explanation is that part of the tracer plume was carried to site 9 by stronger upper level winds. Sites 1 and 10 were located in the creek bed while station 9 was located 80 meters above the bed of Anderson Creek.

(3) The plume elongated considerably in a relatively short travel distance. Elongation factors ranged from 1.8 at 2.5 km to 9.0 at 6.4 km. The elongation factor per kilometer of travel distance was more constant with values ranging from 0.7 to 1.5 indicating an almost linear relationship between the elongation and the travel distance.

One must be careful with the interpretation of the elongation factor since without good wind data at the sampling site one cannot distinguish between a long plume passing by at some speed and a short one passing by at a slower speed. This will be treated in future analysis of this data.

Each of Figures 4 - 6 show the plume passage profiles at a given site for all four experiments. At site 1 (Fig. 4) all profiles are relatively similar in shape. The profile for AC-1 has a larger peak value than the other three. At sites 9 and 10 (Figs. 5 and 6) the profile of AC-1 continues to have the largest peak value. The profiles for AC-2 and AC-3 are

the most similar at these two sites, while that for AC-4 is fairly erratic and not well defined.

From this way of looking at the data, it is implied that at site 1 the transport characteristics of the drainage flow from R1 to site 1 were similar on all four nights. The higher concentration during the first experiment is probably due to the fact the release point R1 was further below the ridge line than in the other three experiments and more tracer was injected in the drainage flow. As stated in the section on integrated samples this result probably reflects the better defined air shed of Anderson Creek between R1 and site 1.

The results of Figures 4 - 6 also indicate that experiments 2 and 3 were conducted in very similar transport conditions in the drainage flow region from R1 to site 10. The first experiment is also very similar, but with higher concentrations, the reason for which have been speculated on above. Although experiment 4 looks similar to the others at site 1, the results at site 9 and 10 indicate a very erratic behavior and perhaps an absence of the drainage flow conditions in the lower part of the valley that existed on the three other nights.

IV. SUMMARY

Four fluorescent particle tracer experiments were performed during the 1979 ASCOT field experiment in Anderson Creek Valley. The results showed considerable interaction between the nocturnal drainage flow fields of Anderson and Putah Creek. Sequential samples taken at three locations in the Anderson Creek air shed measured effective transport velocities for the tracer material of 1 to 2 m/s. The same data also indicated a plume elongation in time that varies approximately linearly with travel distance measured in kilometers. The overall result of the tracer experiments reflects the complexity of the flow fields in the lower part of the Anderson Creek Valley due to various interacting drainage flows.

ACKNOWLEDGEMENTS

Dr. Teymoor Gedayloo and John Archuleta played a major role in data collection and processing. Environmental Systems and Services of Kelseyville, CA provided operational support for the execution of the experiments.

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TABLE I
FP TRACER MATERIAL

<u>Property</u>	<u>Yellow</u>	<u>Green</u>
Type ¹	Lot 1208, #718	Lot 1320, #763
Mass Mean Diameter (μm)	2.74	2.79
Particles per gram	2.31×10^{10}	2.21×10^{10}

¹ Ted Brown Associates, Los Altos Hills, CA.

TABLE II
RELEASE DATA FOR FP EXPERIMENTS

Experiment	Date	<u>Anderson Creek</u> ¹		<u>Putah Creek</u> ²	
		Start	Duration (min)	Start	Duration (min)
FP -1	July 18	2300 ³	25	2300	27
FP -2	July 22	2215	23	2215	25
FP -3	July 24	2225	21	2225	25
FP -4	July 26	2310	21	2310	20

1 500 g of yellow FP, 2.3×10^{10} particles/g; mean mass dia. = $2.7 \mu\text{m}$

2 500 g of green FP, 2.2×10^{10} particles/g; mean mass dia. = $2.8 \mu\text{m}$

3 Pacific Standard Time.

TABLE III

Transport Characteristics for ACV Experiments

Experiment	Sampler ¹	V-eff(m/s)	Elongation Factor	Elongation Factor per Kilometer of Travel
ACV-1	1	1.2	1.8	0.7
ACV-1	9	2.0	1.7	0.7
ACV-1	10	1.0	- -	
ACV-2	1	1.4	1.8	0.7
ACV-2	9	2.0	4.6	0.9
ACV-2	10	1.4	7.0	1.1
ACV-3	1	1.1	1.8	0.7
ACV-3	9	2.0	7.3	1.5
ACV-3	10	1.4	9.0	1.4
ACV-4	1	1.2	2.6	1.0
ACV-4	9	0.7	- -	- -
ACV-4	10	0.6	- -	- -

- ¹ Sampler 1 = 2.5 km from release
 Sampler 9 = 4.9 km from release
 Sampler 10 = 6.4 km from release

**FP RELEASE AND SAMPLING LOCATIONS
ANDERSON CREEK VALLEY
JULY 1979**

Fig. 1. Locations of release points and sampling sites for the FP tracer experiments.

FP - 1, JULY 18-19, 1979

RELEASE START: 2300 PST

Fig. 2a. Integrated FP results for FP-1

FP - 2, JULY 22 - 23, 1979

RELEASE START: 2215 PST

Fig. 2b. Integrated FP results for FP-2

FP - 3, JULY 24-25, 1979
RELEASE START: 2225 PST

Fig. 2c. Integrated FP results for FP-3

FP - 4, JULY 26-27, 1979
RELEASE START: 2310 PST

Fig. 2d. Integrated FP results for FP-4

(a)

(b)

Fig. 3. Tracer Plume concentration as a function of time at three down valley locations. Curves are labeled with kilometers from the tracer release point.

(c)

(d)

Fig. 3. Tracer plume concentration as a function of time at three down valley locations. Curves are labeled with kilometers from the tracer release point.

Fig. 4. Tracer plume concentration as a function of time at 2.5 km from the release point for all experiments.

Fig. 5. Tracer plume concentration as a function of time at 4.9 km from the release point for all experiments.

Fig. 6. Tracer plume concentration as a function of time at 6.4 km from the release point for all experiments.

APPENDIX

LOCATIONS OF FP RELEASE POINTS AND SAMPLERS

<u>Identifier</u>	<u>UTM Coordinates</u> <u>(x,y)</u>	<u>Location</u>
<u>Release Points</u>		
R1	(522.02, 4293.54)	Anderson Creek
R2	(525.70, 4295.01)	Putah Creek
R3	(525.45, 4296.11)	Whispering Pines
<u>Samplers</u>		
1 (I,S) ¹	(524.38, 4292.57)	Socrates Mine Road at Anderson Creek Crossing
2 (I)	(525.73, 4291.61)	Anderson Springs, 1.45 miles from junction of highway 175
3 (I)	(526.71, 4291.57)	Anderson Springs, 0.7 miles from junction of highway 175
4 (I)		Caltrans yard near ES&S-MRI tower
5 (I)	(529.70, 4291.12)	Diamond D Ranch near PNL tower
6 (I)	(527.56, 4292.28)	Highway 175, 0.1 mile south of junction with Socrates Mine Road
7 (I)	(526.45, 4293.52)	Highway 175, 1.3 miles south of Putah Creek release point R2
8 (I)	(527.19, 4292.61)	Highway 175, 0.15 mile north of junction with Socrates Mine Road, entrance to boys camp
9 (I,S)	(526.80, 4292.30)	TV Repeater Site
10 (S)	(528.23, 4291.73)	Highway 175, 0.3 mile south of south turnoff to Anderson Springs

1 - Type of Sampler: I - integrated, S - sequential

Fig. A-1 Locations of FP releases (R), integrated samples (I), and sequential samplers (S).

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A COMPARISON OF OBSERVED AND PREDICTED SLOPE WINDS

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Introduction

Nocturnal drainage winds begin when air adjacent to an inclined surface flows down the slope because it is cooled more than the free air at some distance horizontally from the surface. These slope winds merge and are channeled by the topography to form the larger-scale drainage or mountain winds. This paper discusses the slope flow phase of the drainage wind. The predictions of a simple model for flow down a two-dimensional slope are compared to observations of the drainage wind obtained during the July 1979 ASCOT field study near Middletown, CA.

A Simple Model of Slope Winds

Manins and Sawford (1979a), following Ellison and Turner (1959), have developed a simple model for katabatic flow down a plane surface inclined at an angle α above the horizontal. Coriolis effects are neglected and the ambient air is assumed to be at rest, so that the cross-slope wind is zero and the katabatic flow equations are functions only of time, the down-slope coordinate s and the coordinate n normal to the slope. The air is assumed to be an incompressible, Boussinesq fluid which is in hydrostatic equilibrium along the coordinate n and which has an ambient lapse rate of potential temperature γ . The air adjacent to the slope is then cooled by divergence of the vertical fluxes of radiation and sensible heat, producing a temperature deficit d and a potential temperature distribution:

$$\theta = \theta_0 + \gamma z - d(s, n). \quad (1)$$

The dynamic equations for katabatic flow are further simplified by integration over the normal coordinate n , from the surface to a height H unaffected by the katabatic flow. At this height the wind, the temperature deficit and the Reynolds fluxes of heat and momentum are all assumed to be zero. The resulting equations predict scales of velocity V , depth h and temperature D , defined by:

$$Vh = \int_0^H u \, dn \quad (2a)$$

$$V^2 h = \int_0^H u^2 \, dn \quad (2b)$$

$$VDh = \int_0^H u d \, dn, \quad (2c)$$

which are functions only of time and distance down the slope. The closure problem for the Reynolds fluxes of heat and momentum reduces to specifying the fluxes at the surface. Vertical integration of the dynamic equations does, however, require the additional specification of profile factors S , e.g.,

$$\frac{1}{2} S_1 D h^2 = \int_0^H n d \, dn \quad (2d)$$

$$S_2 D h = \int_0^H d \, dn, \quad (2e)$$

and of the entrainment velocity w_H at the top of the layer. The latter is assumed proportional to the velocity of the layer,

$$w_H = -EV, \quad (3)$$

with the entrainment parameter E an empirical function of the layer Richardson number (Ellison and Turner, 1959),

$$Ri = \frac{g}{\theta_0} \frac{Dh}{V^2} \cos \alpha. \quad (4)$$

The predicted variables V , h and D are good candidates for the parameters of a semi-empirical model of vertical diffusion in drainage flows, i.e. the cross-slope-integrated concentration $\bar{c}^y$ downwind of a continuous source of strength $\dot{Q}$ may be written

$$\bar{c}^y = \frac{\dot{Q}}{Vh} f(s, Ri), \quad (5)$$

where f is a function to be determined. Further, this model is simple enough to permit a variety of sensitivity studies to determine the relative importance of the various terms in the dynamic equations. The major limitations of this model are currently its failure to account for either ambient winds, cross-slope winds in particular, or for horizontal convergence due to inhomogeneity in the cross-slope direction.

An analytical prediction of the slope flow follows from assuming a steady-state, $\gamma = 0$ and neglecting the variation of the profile factors with s ,

$$h = 0.75 Es \quad (6a)$$

$$V^3 = \frac{s}{Ri} \left[\frac{g}{\theta_0} Q \right] \cos \alpha \quad (6b)$$

$$D = Qs/Vh \quad (6c)$$

where $\rho C_p Q$ is the net heat flux out of the drainage layer. The layer Richardson number is found to be independent of s and a function only of the slope angle, a surface drag coefficient C_d , S_1 , S_2 and the functional relationship between Ri and E . Manins and Sawford (1979a) show that the effect of ambient stratification γ is negligible for s less than a scaling distance s_m which decreases with increasing slope angle. Beyond that distance, if $\gamma > 0$ then h is increasingly greater than predicted by (6) and V , D and Ri are smaller.

The authors have extended the analytical solution to account for a change of slope angle by assuming that R_i immediately adjusts to the new slope angle and by conserving either the mass or momentum flux through the slope change. For a decrease in the slope angle this produces an immediate deceleration, thickening and cooling of the drainage layer. Preliminary numerical calculations with the complete equations show that this solution is only qualitatively correct near the slope change, but improves rapidly with distance downslope from the change.

Observations of Slope Winds

Pacific Northwest Laboratory measured profiles of wind speed, wind direction and temperature with a tether sonde located at the PDC-1 site, at an elevation of 850 m MSL on the south-southeast slope of Cobb Mountain. The summit of Cobb Mountain is 1440 m, and the slope angle is roughly 22° down to an elevation of 900 m and 6.5° thereafter.

The best drainage wind observations were made on the night of July 24-25, 1979, particularly after midnight PST. Values of h , V , D , S_1 , S_2 and R_i were calculated from the data for seven profiles made from 8 p.m. July 24 to 5 a.m. July 25. The down-slope direction was defined for each profile by the wind direction at the height of the maximum wind speed, and the wind data were resolved into down-slope and cross-slope components. For most of the profiles this direction was south-southeast, the direction of a line between the top of Cobb Mountain and PDC-1. Near ground-level the wind direction was west, corresponding to the slope in the immediate vicinity of PDC-1.

Table 1 shows the calculated values for three of the seven profiles. These profiles were selected on the basis of the conformity of the wind profiles to the assumptions of the model.

Table 1. Slope Wind Observations at PDC-1, July 25, 1979.
 Model Predictions Assume $C_d = .047$, (a) $\rho C_p Q = 115 \text{ watts/m}^2$,
 and (b) $\rho C_p Q = 40 \text{ watts/m}^2$

<u>t</u> <u>(PST)</u>	<u>h</u> <u>(m)</u>	<u>V</u> <u>(m/s)</u>	<u>D</u> <u>(°C)</u>	<u>S₁</u>	<u>S₂</u>	<u>Ri</u>
0011	60	1.92	1.19	0.93	1.46	0.62
0309	72	2.35	1.40	0.98	1.56	0.55
0500	67	2.14	1.01	1.06	1.48	0.47
Average	66	2.14	1.20	0.99	1.50	0.55
Model						
(a)	36	2.14	2.23	0.5	0.9	0.55
(b)	36	1.51	1.10	0.5	0.9	0.55

They have down-slope winds with maxima of 3 to 4 m/s, which decrease above the drainage layer to 1 m/s or less, and cross-slope winds below 1 m/s. Profiles of wind and temperature for the 0309 sounding are shown in Figs. 1 and 2. The Ri for these profiles is also consistently on the order of 0.5, while those for the excluded profiles are greater than one. The lower values are more consistent with the data of Ellison and Turner (1959) and the calculations of Manins and Sawford (1979a).

Although the observations at the PDC-1 site do not satisfy the analytical model assumptions of cross-slope homogeneity and $\gamma=0$, Table 1 also lists two predictions of the simple slope flow model. Since C_d and Q were not directly measured, values for these parameters have been chosen to match the predictions to the observations. The selected parameter values and the resultant predictions are discussed below, along with the probable reasons for the existing discrepancies between the analytical slope flow model and the observations.

Figure 1. Drainage wind profiles at PDC-1, 0309-0316 PST, July 25, 1979

Figure 2. Potential temperature profile at PDC-1, 0309-0316 PST, July 25, 1979

A drag coefficient C_d equal to 0.05 is necessary for $Ri = 0.55$ on a 6.5° slope. This value is quite high, particularly for stable stratification, but is consistent with Yamada's inclusion of canopy drag to match his modeled winds to the observations. With $C_d = 0.05$ the predicted depth of the slope flow is only about half of that observed. The model accounts for the vertical flow convergence due to the slope change, but there may also be some horizontal convergence since the terrain is not homogeneous in the cross-slope direction. A more likely explanation, however, is adiabatic warming of the drainage flow.

Manins and Sawford (1979a) find that the analytical solution with $\gamma = 0$ is a good approximation on a 5° slope if s is less than 4 km. For a 22° slope, however, this limit is only about 500 m, less than one-third the distance from the top of Cobb Mountain to PDC-1. Hence the ambient stratification needs to be included in the slope flow model. The resultant warming of the drainage layer relative to the ambient air will decelerate and thicken the flow, as observed, but the overall result is not possible to predict analytically. A numerical solution is currently being developed for the case of a change in slope with $\gamma \neq 0$.

A net heat flux of 115 watts/m^2 is necessary for the analytical model to reproduce the observed wind. This is two to three times the surface net radiation and thus is quite large. Again horizontal convergence could account for this discrepancy (Manins and Sawford, 1979b). With this heat flux and a predicted h only half of that observed, D is predicted to be twice that observed. A more realistic heat flux of 40 watts/m^2 gives a predicted slope wind of 1.5 m/s and a bulk temperature deficit of 1.1°C .

Finally, S_1 and S_2 in Table 1 are quite consistent at values of 1 and 1.5. Values close to these were found for most of the July 24-25 soundings. Ellison and Turner (1959) found values of 0.25 and

0.75 for laboratory-scale flows and Manins and Sawford (1979a) report atmospheric values of 0.5 and 0.9. The latter values were used in the preceding model calculations since the larger values at PDC-1 may be associated with the slope change which is about 300 m upslope of the measurement site and may not be applicable to most of the slope.

Conclusion

Drainage wind observations at the PDC-1 site have been compared to a simple slope flow model. Considering that the model assumptions of cross-slope homogeneity and a neutral ambient atmosphere are not satisfied by the observations, this initial comparison is encouraging. Two approaches are being taken for improvement. First a numerical solution is being developed to include the effects of slope change and ambient stratification. The results of Manins and Sawford (1979a) suggest that the numerical predictions will more closely match the observations of Table 1. Second, more detailed slope flow measurements are being made near Richland, WA., at a site which more closely approximates the two-dimensional assumptions of the model. Three towers, instrumented to measure wind and temperature, and a tethered sonde will be used to define the growth of the drainage layer, including its alteration by a steep-to-shallow slope change.

Acknowledgements

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TWO-DIMENSIONAL SLOPE WIND SIMULATIONS IN THE FINITE ELEMENT APPROXIMATION

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ABSTRACT

The hydrostatic fluid dynamics model developed at LLL has been used to simulate the development of katabatic winds. This model solves the Navier-Stokes equations in the Boussinesq approximation by the finite element method. Preliminary results indicate that to obtain physically reasonable results one has to choose unequal diffusion parameters in the horizontal (K_x) and vertical (K_z). The maximum velocities obtained with $K_z = 1 \text{ m}^2/\text{sec}$ and $K_x = 100 \text{ m}^2/\text{sec}$ are of the order of 2.5 m/sec for a slope of 0.2. Profiles of the downslope velocities will be presented at different points in the flow. As expected, the magnitude of the vertical diffusion coefficient K_z controls the depth of the flow which seems to increase only slightly with downhill distance, and the magnitude of the flow increases with cooling rate and slope.

INTRODUCTION

Slope winds, i.e., winds produced by the sustained cooling of a slope at night seem to be the most important ventilation mechanism in complex terrain in the

absence of strong synoptic pressure gradients, and thus become important for air quality models in complex terrain.

The historical development of slope winds modeling and observation are treated in Thyer (1966) and more recently in Manins and Sawford (1979a). The first model of slope winds was proposed by Prandtl (1942, see Sutton 1953) and was limited by the assumption that the winds were steady along the slope, thus reducing the problem to one dimension. Another approach is to consider only an average flow within the cooled layer, eliminating all internal structure of the wind field (Petkovsek and Hocevar, 1971). This approach has been generalized by Manins and Sawford who impose three profile factors on the flow, thus the vertical structure of the katabatic wind field is determined a priori and the subsequent time development of the flow is dependent on the initial guess of these profile factors.

The third approach to the modeling of drainage is the numerical solution of a set of primitive equations for the atmosphere. This was first done by Thyer (1966), L. M. Leslie and K. R. Smith (1974) and more recently by C. P. Stevens (1979). The latter calculation was only done for one set of parameters, i.e., one slope and one cooling rate and with closed boundaries. Yamada (1980) has applied his second order closure model to the problem of slope winds.

In the following, I will present results for a primitive equation model (Navier-Stokes equations in the Boussinesq approximation) solved by a finite element numerical algorithm. In the next section I will present a short summary of the equations. The discussion of the result will emphasize variation of the key parameters: diffusivity, slope and cooling rates. Finally, the conclusions drawn from these runs and the feasibility of an extension of the model to three dimensions will be discussed.

MODEL EQUATIONS

The model used was the LLL hydrostatic fluid dynamics model developed by Chan et al. (1979). This model uses the Navier-Stokes equations in the Boussinesq approximation i.e., in two dimensions

$$\frac{\partial u}{\partial t} + u \frac{\partial u}{\partial x} + w \frac{\partial u}{\partial z} = - \frac{\partial p}{\partial x} + K_x \frac{\partial^2 u}{\partial x^2} + K_y \frac{\partial^2 u}{\partial y^2} \quad (1)$$

$$\frac{\partial p}{\partial y} = \rho g \gamma T \quad (2)$$

$$\frac{\partial u}{\partial x} + \frac{\partial w}{\partial z} = 0 \quad (3)$$

$$\frac{\partial T}{\partial t} + u \frac{\partial T}{\partial x} + w \frac{\partial T}{\partial z} = K_x^T \frac{\partial^2 T}{\partial x^2} + K_z^T \frac{\partial^2 T}{\partial z^2} \quad (3)$$

Where u is the horizontal velocity, w is the vertical velocity, p the pressure deviation from equilibrium, T is the temperature deviation from equilibrium, ρ is the density, g the acceleration of gravity, γ the coefficient of thermal expansion K_x , K_z , K_x^T and K_z^T are the momentum and temperature diffusion coefficients respectively.

To solve this system of equations in the finite element method, one expands the unknowns, u , w , p , and T in the equation in highly localized polynomials and solves for the expansion coefficients, i.e.,

$$\begin{aligned} u &= \sum_i u_i \phi_i(x, z) \\ w &= \sum_i w_i \phi_i(x, z) \\ T &= \sum_i T_i \phi_i(x, z) \\ p &= \sum_i p_i \psi_i(x, z) \end{aligned} \quad (5)$$

These expansions are then substituted into the original equations and after applying some weighted residual method (in this case, Galerkin for equations (1) and (4), least squares for equations (2) and (3)), one obtains a set of algebraic equations which may be solved for the expansion coefficients. For a much more complete discussion of the numerical methods employed, the reader is referred to Chan (1979).

The choice of the hydrostatic code for these calculations deserves some discussion. The non-hydrostatic FEM code has at the moment the fundamental limitation of requiring the pressure basis functions to be polynomials of one order lower than the basis functions for the velocity, and temperature. This means in practice that the pressure basis functions are restricted to be either piecewise constant or piecewise linear polynomials. On the other hand, a linear temperature distribution, the simplest non-trivial temperature deviation from equilibrium, will induce a quadratic pressure variation. Trying to represent this quadratic pressure fluctuation with either linear or piecewise constant pressure basis functions leads to a large error in the pressure gradient across the grid in the presence of topography (Tuerpe, 1980). The hydrostatic fluid dynamics code has the capability of quadratic pressure basis functions and has been used in the obtaining the results of the next section.

RESULTS

Most of the runs, with the exception of those with a different slope, were made with the grid of Fig. 1, which has 341 node points, 11 in the vertical and 31 in the horizontal. The basis functions for all variables are quadratic, defined on the standard 9-node element. The boundary conditions are as indicated in Fig. 1, with open top and lateral boundary, and a no flow bottom boundary. The bottom boundary was then cooled uniformly according to

$$T_G = C_1 \sin \frac{t}{C_2}$$

where T_G = ground temperature, and

t = time.

Figure 2 shows the vector field of a typical run after about 2 hrs. A drainage layer about 100 meters high is formed above the ground and the air starts flowing down the mountain with a velocity of about 2-3 m/sec. In all of these runs the momentum diffusivities are assumed to be equal to the temperature diffusivities.

Figures 3, 4 and 5 show the profiles of the velocity along the slope at different points on the slope. It is seen that in all three cases the depth of the drainage layer increases slightly with distance along the slope and that this depth is controlled by the magnitude of K_z , i.e., the larger K_z , the larger the drainage layer. Fig. 6 shows the maximum velocity within the drainage layer as a function of slope. As expected, for the same rate of cooling of the ground, the larger the slope, the larger the drainage velocity. Fig. 7 shows the maximum velocity along the slope as a function of the vertical diffusivity. The falloff in the magnitude of the velocity near the bottom of the slope occurs because the drainage layer abruptly widens due to the change in slope at the bottom and this effect is more pronounced for a wider initial drainage layer, i.e., a run with a higher K_z . Fig. 8 shows the maximum velocity as a function of various parameters. Curve 2 has the same parameters as curve 1, except plotted at a later time. Since equilibrium has not yet been reached, the downslope velocity is still increasing, as expected. Curve 3 on Fig. 8 is the same as curve 1 except with the cooling rate doubled.

Finally, Fig. 9 shows an attempt to model upslope flow. In this run the surface is heated, as occurs early in the morning when the sun first begins to shine on the slope. This situation is much more unstable than that in which the slope was

cooled. Here cool air is above warm air and the slope changes induce instability into the flow which produce large vertical velocities. It is not quite clear whether these large vertical velocities in the model run come about because of the nature of the hydrostatic approximation, which breaks down at points of high vertical acceleration, or whether they are a realistic consequence of the physics of this highly unstable situation.

CONCLUSIONS

The results reported in this paper have demonstrated that the finite element approximation is useful for solutions of the Navier-Stokes equations with atmospheric applications, provided one has chosen quadratic or higher order polynomials as the pressure basis functions when atmospheric process over complex terrain are modeled. Detailed comparison of these two-dimensional results to experiment, other than very qualitative observations, are not appropriate, since two-dimensional slopes do not usually occur in nature. However, the orders of magnitude of the velocity (2-4 m/sec) compares well with the results of the 1979 ASCOT Geysers experiment and the measurements of Manins and Sawford (1979b). The vertical profile of the drainage velocity is very similar to that measured by Manins and Sawford, but differs with the 1979 Geysers data where the profile is much broader than in the present calculations. This might be due to a canopy drag factor, neglected in the present model. Yamada (1980) appears to obtain broader profiles when such a canopy drag factor is included. In addition, the terrain in the Geysers is very complex, while the site for the Manins and Sawford experiment was close to an idealized two-dimensional slope.

The calculations reported here were done as an intermediate step to the development of a three-dimensional atmospheric dynamics code valid in complex

terrain. One should also investigate the validity of the hydrostatic approximation for slope winds. Here the requirement of a quadratic pressure basis function in the finite element approximation causes some difficulty, since the velocity and temperature basis functions should be one order higher than the pressure basis functions for a nonhydrostatic formulation. This would mean that the basis functions for the velocity and temperature would have to be cubic, however, mesh generators for grids appropriate to cubic basis functions do not exist. One approach to overcome this would be to construct an "almost" nonhydrostatic code by reinserting the vertical acceleration terms but not the diffusion terms into the hydrostatic formulation. This procedure also allows one to retain the simple boundary conditions of the hydrostatic code. Another approach would simply be to ignore the bad pressure solution obtained with a nonhydrostatic fluid dynamics code where the pressure and velocity basis functions are of the same order. Here the spurious part of the pressure solution does not act as a driving term for the velocity.

Both the development of the 3-D hydrostatic fluid dynamics code and the reformulation of the nonhydrostatic code as a check on the validity of the hydrostatic approximation are being implemented at LLL.

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FIGURE 1. The grid used in the model runs for a slope of .2. Boundary conditions are indicated on the figure.

FIGURE 2. A typical vector flow-field at $t = 6800$ sec, $K_x = K_x^T = 100$ m^2/sec , $K_z = K_z^T = 1$ m^2/sec , $T_G = -20 \sin t/27000$, slope = 0.2.

M-10

FIGURE 3. Profile of the velocity along the slope, same parameters as figure 2 except for $K_z^T = K_z = 0.1 \text{ m}^2/\text{sec}$ and $t = 4153 \text{ sec}$.

FIGURE 4. Profile of the velocity along the slope, same parameters as Fig. 3 except $K_z = K_z^T = 1.0 \text{ m}^2/\text{sec}$.

FIGURE 5. Profile of velocity along the slope, same parameters as Fig. 3 except $K_z = K_z^T = 5.0 \text{ m}^2/\text{sec}$.

FIGURE 6. Maximum velocity along the slope as a function of K_z . $T_G = -20$
 $\sin t/27000$, $t = 4100$ sec, $K_x = K_x^T = 100$ m²/sec, slope = .2

Curve 1: $K_z = K_z^T = 0.1$ m²/sec.

Curve 2: $K_z = K_z^T = 1.0$ m²/sec.

Curve 3: $K_z = K_z^T = 5.0$ m²/sec.

FIGURE 7. Maximum velocity along the slope as a function of slope. $T_G = -20$
 $\sin t/27000$, $t = 4100$ sec, $K_x = K_x^T = 100$ m²/sec $K_z =$
 $K_z^T = 1.0$

Curve 1: Slope = 0.1
 Curve 2: Slope = 0.2
 Curve 3: Slope = 0.3

FIGURE 8. Maximum velocity along the slope as a function of time and cooling rate. $K_x = K_x^T = 100 \text{ m}^2/\text{sec}$, $K_z = K_z^T = 1 \text{ m}^2/\text{sec}$, slope = 0.2

Curve 1: $T_G = -20 \sin t/2700$, $t = 4058 \text{ sec}$

Curve 2: $T_G = -20 \sin t/2700$, $t = 6800 \text{ sec}$

Curve 3: $T_G = -40 \sin t/2700$, $t = 4092 \text{ sec}$.

FIGURE 9. Upslope wind after 2700 sec. $T_G = 20 \sin t/2700$, $K_x = K_x^T = 100 \text{ m}^2/\text{sec}$, $K_z = K_z^T = 1 \text{ m}^2/\text{sec}$.

MASS-CONSISTENT MODELING, REVISITED

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Introduction

Recent amendments of the Clean Air Act require the use of air quality models to determine impacts of pollutant emissions from new and existing facilities. Because of the current need for greater production of domestic liquid fuel, many new energy facilities will be located in the complex terrain regions of the energy-rich West. Thus, there is a continued and expanded interest in improving all phases of air quality modeling for complex terrain.

A generalized air quality model consists of many components or modules. These modules account for the following phenomena:

- Topographic and surface effects.
- Size, geometry and location of pollutant sources.
- Mean meteorological variables of wind speed and direction, solar radiation, temperature, pressure, density and phases of water.
- Airborne evolution of gaseous and particulate pollution.
- Wet and dry removal of pollutants.
- Effects of air pollutant concentrations and quantities of deposited pollutants on visibility degradation and acidification of soils and water

The proper performance of such an air quality model requires, among other elements, a good description of the mean wind fields over the complex terrain under consideration. Hence, the subject of the present paper is to reconsider the manner in which mass-conserving wind fields are derived for airflow over complex terrain. The current methods for deriving mass-consistent winds are analyzed; a mass-consistent interpolation scheme is proposed; and a set of analytically-derived wind fields is proposed for checking interpolation schemes and wind field simulations.

Computational Domain

Figure 1 shows a Cartesian grid (x, y, z) fixed over a region of complex terrain, indicated by $\bar{z}_s(x, y)$. The upper layer, denoted by $\bar{H}(x, y)$, may be an inversion surface, the tropopause, an "imaginary" horizontal surface, or some other prescribed surface or layer. In general, $\bar{H}$ is a function of time t , while $\bar{z}_s$ is fixed over the periods of time important for air quality modeling. However, if $\bar{z}_s$ represents an extensive water surface, then $\bar{z}_s$ may vary with time because of wave motion. The horizontal dimensions of the domain, denoted by L_x and L_y , are fixed quantities. Usually, we assume that

$$L_x \text{ and } L_y \gg \bar{h}, \quad (1)$$

where $\bar{h}$ is defined by

$$\bar{h}(x, y) = \bar{H}(x, y) - \bar{z}_s(x, y). \quad (2)$$

Mean State of the Atmosphere

The mean state of the atmosphere in the domain of Figure 1 is defined by the meteorological variables:

- Velocity field, $\bar{V} = (\bar{u}, \bar{v}, \bar{w})$.
- Density field, $\bar{\rho}$.
- Pressure field, $\bar{p}$.
- Temperature field, $\bar{T}$.
- Phases of water, $\bar{r}_i$.
- Solar radiation as a function of band widths, $\bar{R}_i$.
- Correlations between mean fields, $\bar{c}_{ij}$.

These quantities are governed by a set of Reynolds-averaged equations which account for the conservation of mass, momentum, energy and water phases, atmospheric and surface radiation balances, and the thermodynamic state of the air. If the system parameters, and the initial and boundary conditions are properly specified over the domain, then a set of unique solutions exists for the equations.

THE CARTESIAN GRID

Figure 1. Schematic of a Cartesian grid over a region of complex terrain.

The current paper is concerned with the unique, mean, velocity field

$$\bar{V}_* = (\bar{u}_*, \bar{v}_*, \bar{w}_*). \quad (3)$$

Mass-Consistent Modeling

Mass-consistent modeling is an attempt to derive "snapshots" of the mean velocity field $\bar{V}_*$ at a set of discrete times $t_0, t_1, t_2, \dots, t_p$. The usual starting point of mass-consistent modeling is a discrete set of spatial and temporal wind measurements taken at various points in and around the domain in Figure 1. This set of quantities may also be derived by large-scale model simulations. Suppose this set of winds is given by

$$\bar{V}_m = (\bar{u}_m, \bar{v}_m, \bar{w}_m). \quad (4)$$

However, for measured winds, $\bar{w}_m$ is rarely given.

The next step in the process is the application of a temporal-spatial, interpolation-extrapolation scheme to the field $\bar{V}_m$, resulting in a continuous velocity field in space and time:

$$\bar{V}_0 = (\bar{u}_0, \bar{v}_0, \bar{w}_0). \quad (5)$$

Finally, this field $\bar{V}_0$ is converted into a mass-consistent flow field

$$\bar{V}_1 = (\bar{u}_1, \bar{v}_1, \bar{w}_1). \quad (6)$$

for the times $t_0, t_1, t_2, \dots, t_p$. Mass consistency is guaranteed by having $\bar{V}_1$ satisfy the continuity equation

$$\frac{d}{dt} [\ln \bar{\rho}] + \bar{u}_{,x} + \bar{v}_{,y} + \bar{w}_{,z} = 0 \quad (7)$$

over the time interval from t_0 to t_p .

Relationship of $\bar{V}_1$ to $\bar{V}_*$

A solution of Eq. (7) over the domain in Figure 1 is not unique unless further restrictions are placed on it. In mass-consistent modeling a further restriction is often some "goodness-of-fit" criterion between $\bar{V}_1$ and $\bar{V}_0$. That is, a norm is defined

$$||\bar{V}_1 - \bar{V}_0||$$

and some minimization criterion is imposed. For example, Sherman [1] used an integral norm and the criterion of a least-squares fit of $\bar{V}_1$ to $\bar{V}_0$. In this technique the derivation of a unique $\bar{V}_1$ requires a solution of a Poisson equation for a Lagrangian multiplier over the domain of computation. Huang and Drake [2] derived a finite-difference, iterative scheme that converges to a unique field $\bar{V}_1$; however, it is not clear what the "goodness-of-fit" of $\bar{V}_1$ to $\bar{V}_0$ is in this procedure.

An investigator should always keep in mind that even though $\bar{V}_1$ may be, in some sense, close to $\bar{V}_0$, it is unknown how close $\bar{V}_1$ is to the desired field $\bar{V}$. Other solutions of Eq. (7) may even be closer to $\bar{V}_0$ and/or $\bar{V}_*$ than $\bar{V}_1$, for example

$$\bar{V}_a = (a \bar{u}_1, a \bar{v}_1, a \bar{w}_1), \quad (8)$$

where "a" is an arbitrary real number whose magnitude $|a|$ is restricted in some fashion. Still other solutions of Eq. (7) can be easily derived from $\bar{V}_0$ and Eq. (7), such as:

$$\bar{V}_2 = (\bar{u}_2, \bar{v}_2, \bar{w}_2) = (\bar{u}_0, \bar{v}_0, \bar{w}_2), \quad (9)$$

$$\bar{V}_3 = (\bar{u}_3, \bar{v}_3, \bar{w}_3) = (\bar{u}_0, \bar{v}_3, \bar{w}_0), \quad (10)$$

$$\bar{V}_4 = (\bar{u}_4, \bar{v}_4, \bar{w}_4) = (\bar{u}_4, \bar{v}_0, \bar{w}_0), \quad (11)$$

where $\bar{w}_2, \bar{v}_3, \bar{u}_4$ are respectively derived from Eq. (7). The linear combination of the $\bar{V}_i$'s,

$$\bar{V}_5 = b \bar{V}_1 + c \bar{V}_2 + d \bar{V}_3 + e \bar{V}_4, \quad (12)$$

where b, c, d and e are arbitrary real numbers, gives another infinite set of solutions that may be close to $\bar{V}_0$ and/or $\bar{V}$. The venn diagram in Figure 2 indicates the "close" relationship between* the solutions of Eq. (7) and the interpolated-extrapolated field $\bar{V}_0$, and the "more distant" relationship with $\bar{V}$. By construction, it is assumed that the solutions given by Eq. (6), (8) to (12) are in some neighborhood of $\bar{V}_0$ (a neighborhood defined by distance d_n), while the distance $||\bar{V}_* - \bar{V}_0||$ between $\bar{V}_0$ and $\bar{V}_*$ is basically unknown.

Hence, the burden of finding the "best" $\bar{V}_1$ compared to $\bar{V}$ is on the interpolation-extrapolation scheme that produces $\bar{V}_0$, not on the mass-conservation scheme. For maximum efficiency and accuracy we should apply the best

MEAN VELOCITY FIELD VENN DIAGRAM

Figure 2. A venn diagram showing the relationships between the desired mean wind field $\bar{V}_*$, the interpolated-extrapolated field $\bar{V}_0$, and an infinite set of mass-consistent fields, $\bar{V}_i$, $i = a, 1, 2, 3, 4, 5$.

possible interpolation-extrapolation scheme to $\bar{V}_m$ and then use the "cheapest" possible mass-consistent scheme.

Anelastic Atmospheres

The anelastic assumption of Ogura and Phillips [3] reduces the first term in Eq. (7) to

$$\frac{d [\ell n \bar{\rho}]}{dt} \approx \frac{d [\ell n \bar{\rho}]}{dz} \bar{w}, \quad (13)$$

where $\bar{\rho} = \bar{\rho}(z)$. This approximation eliminates the propagation of sound waves and is valid for mesoscale flow problems. If the depth of the domain $\bar{h}$ is less than 1 to 3 km, $\bar{\rho}(z)$ can be replaced by a constant. Thus, we have an incompressible, nonstratified flow governed by

$$\bar{u}_{,x} + \bar{v}_{,y} + \bar{w}_{,z} = 0, \quad (14)$$

while an incompressible, density-stratified flow is governed by

$$\bar{u}_{,x} + \bar{v}_{,y} + \bar{w}_{,z} = - \frac{d [\ell n \bar{\rho}]}{dz} \bar{w}. \quad (15)$$

Perturbed, Dry Adiabatic Atmospheres

For most mesoscale, air quality applications, the atmosphere is a perturbation from a dry adiabatic atmosphere. That is,

$$\theta = \text{potential temperature} = \Theta [1 + f(z)], \quad (16)$$

where Θ is the potential temperature of the dry adiabat and $f(z)$ is a perturbation referred to sea level,

$$|f(z)| \ll 1, f(0) = 0. \quad (17)$$

The state of this perturbed atmosphere is given by the hydrostatic law, Eq. (16) and (17), and the gas law:

$$\bar{\pi} = 1 - \frac{1}{H_a} \int_0^z \frac{dz}{1 + f(z)}, \quad (18)$$

$$\bar{p} = \bar{p}(0) [\bar{\pi}]^{7/2} \quad (19)$$

$$\bar{\rho} = \bar{\rho}(0) [\bar{\pi}]^{5/2} / [1 + f(z)], \quad (20)$$

$$\bar{T} = \Theta [1 + f(z)] \bar{\pi}. \quad (21)$$

The parameters in Eq. (18) to (20) are given by

$$\begin{aligned} H_a &= \text{height of the adiabatic atmosphere} = \frac{c_p \Theta}{g} \\ &= 102.4 \Theta, \end{aligned} \quad (22)$$

$$\bar{p}(0) = 101,300 \text{ newtons/m}^2, \quad (23)$$

$$\bar{\rho}(0) = 353.1/\Theta \text{ kg/m}^3, \quad (24)$$

where the specific heat $c_p = 1004.5 \text{ m}^2/\text{sec}^2 - ^\circ\text{K}$ and $g = 9.81 \text{ m/sec}^2$. Thus, height H_a is approximately equal to 30 km (well above the tropopause).

From Eq. (16) and (20) we see that a 30% variation in atmospheric thermal structure (from unstable conditions to stable conditions) will result in less than a 10% variation in density. Hence, for mass-consistent modeling on the mesoscale, the dry-adiabatic quantity for $\bar{p}(z)$ (i.e., $f(z) = 0$ for all (z)) is sufficient. For this case, $\bar{p}$ is given by Eq. (22) and (24), and

$$\bar{p}(z) = \bar{p}(0) \left[1 - \frac{z}{H_a}\right]^{5/2}. \quad (25)$$

Terrain-Following Coordinate System

In order to improve the interpolation and interpretation of measured wind fields, the Cartesian domain in Figure 1 is transformed to the unit cube shown in Figure 3. The (α, β, γ) coordinates form a terrain-following system defined by

$$x = L_x \alpha, \quad (26)$$

$$y = L_y \beta, \quad (27)$$

$$z = \bar{z}_s + \gamma \bar{h} = z_s + \gamma h. \quad (28)$$

The parameters and dependent variables transform in the following way:

$$\bar{H}(x, y) = H(\alpha, \beta), \quad \bar{z}_s(x, y) = z_s(\alpha, \beta), \quad (29)$$

$$\bar{h}(x, y) = h(\alpha, \beta) = H(\alpha, \beta) - z_s(\alpha, \beta), \quad (30)$$

THE TRANSFORMED GRID

Figure 3. The image of the domain in Figure 1 under the terrain-following coordinate transformation defined by Equations (26) to (28).

$$\bar{\rho}(z) = \rho(\alpha, \beta, \gamma), \quad \bar{u}(x, y, z) = u(\alpha, \beta, \gamma). \quad (31)$$

$$\bar{v}(x, y, z) = v(\alpha, \beta, \gamma), \quad \bar{w}(x, y, z) = w(\alpha, \beta, \gamma). \quad (32)$$

Writing Eq. (15) in terms of Eq. (26) to (32) gives

$$\begin{aligned} (u h)_{,\alpha} + \left(\frac{L_x}{L_y} v h\right)_{,\beta} + (L_x w - \dot{z}_s - \gamma \dot{h})_{,\gamma} &= \\ &= -L_x \frac{\partial (\ln \rho)}{\partial \gamma} w, \end{aligned} \quad (33)$$

where

$$\dot{z}_s = z_{s,\alpha} u + \frac{L_x}{L_y} z_{s,\beta} v, \quad (34)$$

$$\dot{h} = h_{,\alpha} u + \frac{L_x}{L_y} h_{,\beta} v. \quad (35)$$

If (u, v, w) are replaced by

$$U = \rho u h, \quad (36)$$

$$V = \frac{L_x}{L_y} \rho v h, \quad (37)$$

$$W = \rho L_x w - \rho \dot{z}_s - \rho \gamma \dot{h}, \quad (38)$$

then Eq. (33) becomes

$$\boxed{U_{,\alpha} + V_{,\beta} + W_{,\gamma} = AU + BV}, \quad (39)$$

where

$$A = (\ln \rho)_{,\alpha} - (\ln \rho)_{,\gamma} \left[\frac{z_{s,\alpha}}{h} + \gamma (\ln h)_{,\alpha} \right], \quad (40)$$

$$B = (\ln \rho)_{,\beta} - (\ln \rho)_{,\gamma} \left[\frac{z_{s,\beta}}{h} + \gamma (\ln h)_{,\beta} \right]. \quad (41)$$

Interpretation of Equation 39

Integrating Eq. 39 with respect to γ gives

$$W = \int_0^Y [U, V] d\gamma, \quad (42)$$

where $W(\alpha, \beta, 0) = 0$ and

$$[U, V] = AU + BV - U_{,\alpha} - V_{,\beta}. \quad (43)$$

A Mass-Consistent Wind Field

Suppose $\underline{y}_0 = (U_0, V_0, W_0)$ is the transformed interpolated-extrapolated wind field in Eq. (5). Because $\bar{w}_0$ is usually not measured, W_0 may be unknown. A mass-consistent wind field corresponding to $\underline{y}_0$ can be given by

$$\underline{y}_1 = (U_1, V_1, W_1) = (U_0, V_0, W_1), \quad (44)$$

where W_1 is defined by

$$W_1 = \int_0^Y [U_0, V_0] d\gamma. \quad (45)$$

If W_0 is known, then $|W_1 - W_0|$ will be a measure of how good the interpolated and/or measured winds are.

W(α, β, γ) is Zero

If W is identically zero over the unit cube, then

$$w = \frac{1}{L_x} [z_s^2 + \gamma^2]. \quad (46)$$

That is, the winds follow the terrain near the bottom of the domain, follow the upper surface near the top of the domain, and follow transitional paths in between these extremes.

ρ is Constant

If ρ is constant, the flow is non-stratified and A and B are identically zero. If A and B are identically zero, ρ need not be constant; however, for atmospheric winds, $A = B = 0$ usually implies $\rho = \text{constant}$. Hence Eq. (39) reduces to

$$U_{,\alpha} + V_{,\beta} + W_{,\gamma} = 0. \quad (47)$$

No Stratification, General Lifting or Subsidence, and Veering or Backing

If the flow is density stratified, then

$$BU + CV \neq 0. \quad (48)$$

If there is general lifting or subsidence over the region in which the domain of computation is embedded, then $W \neq 0$ and

$$U_{,\alpha} + V_{,\beta} \neq 0. \quad (49)$$

If there is veering or backing of the winds, the divergence of (U, V) is nonzero and $W \neq 0$.

Conversely, if there is no stratification, no regional lifting or subsidence, and no veering or backing of the winds, then Eq. (39) reduces to

$$\boxed{U_{,\alpha} + V_{,\beta} = 0, W = 0}. \quad (50)$$

The general solution of Eq. (50) can be written as

$$U = \sum_n a_n(\gamma) b_n(\alpha, \beta), \quad (51)$$

$$V = \sum_n a_n(\gamma) d_n(\alpha, \beta), \quad (52)$$

where $i = \sqrt{-1}$ and

$$F_n(\alpha + i\beta) = b_n(\alpha, \beta) - i d_n(\alpha, \beta) \quad (53)$$

is an arbitrary, analytic function of the complex variable $\alpha + i\beta$. The winds given by Eq. (51) and (52) are mass-consistent. The vertical components of these winds are only due to the forcing of the terrain and the upper surface $\bar{H}$.

Analytically-Derived Wind Fields

The efficiency and accuracy of interpolation schemes and wind field simulations can be easily checked if a set of analytically-defined wind fields exists. These fields can be derived by choosing specific forms for U and V and then evaluating W from Eq. (42).

The Transformed Wind Fields

The general forms of U and V are given by

$$U = a f_1(\gamma) [f_2(\gamma) f_3(\alpha, \beta) + f_4(\alpha, \beta)], \quad (54)$$

$$V = b f_1(\gamma) [f_5(\gamma) f_6(\alpha, \beta) + f_7(\alpha, \beta)], \quad (55)$$

where a and b are scale factors, $f_1(\gamma)$ accounts for the vertical shear in the surface and boundary layers, $f_2(\gamma)$ and $f_5(\gamma)$ account for veering or backing, $f_3(\alpha, \beta)$ and $f_6(\alpha, \beta)$ account for the major horizontal components of the wind, and $f_4(\alpha, \beta)$ and $f_7(\alpha, \beta)$ account for general lifting and subsidence of the regional atmosphere. In addition, we assume that

$$f_{3,\alpha} = -f_{6,\beta}, \quad f_{4,\alpha} \neq -f_{7,\beta}, \quad (56)$$

$$|f_2 f_3| \gg |f_4|, \quad |f_5 f_6| \gg |f_7|. \quad (57)$$

The corresponding expression for W is given by

$$\begin{aligned} W = & a f_3(\alpha, \beta) f_8(\alpha, \beta, \gamma) + a f_4(\alpha, \beta) f_9(\alpha, \beta, \gamma) \quad (58) \\ & + b f_6(\alpha, \beta) f_{10}(\alpha, \beta, \gamma) + b f_7(\alpha, \beta) f_{11}(\alpha, \beta, \gamma) \\ & + f_{3,\alpha}(\alpha, \beta) f_{12}(\gamma) - [a f_{4,\beta}(\alpha, \beta) + b f_{7,\beta}(\alpha, \beta)] f_{13}(\gamma), \end{aligned}$$

where

$$f_8 = \int_0^\gamma A f_1 f_2 d\gamma, \quad f_9 = \int_0^\gamma A f_1 d\gamma, \quad (59)$$

$$f_{10} = \int_0^\gamma B f_1 f_5 d\gamma, \quad f_{11} = \int_0^\gamma B f_1 d\gamma, \quad (60)$$

$$f_{12} = b \int_0^\gamma f_1 f_5 d\gamma - a \int_0^\gamma f_1 f_2 d\gamma, \quad (61)$$

$$f_{13} = \int_0^\gamma f_1 d\gamma. \quad (62)$$

If there is no density stratification, then $f_8 = f_9 = f_{10} = f_{11} = 0$, and W reduces to a rather simple formula. Further, if there is no veering or backing of the wind, then $f_2 = f_5 = 1$, and W becomes

$$\begin{aligned} W = & [(b - a) f_{3,\alpha}(\alpha, \beta) - a f_{4,\beta}(\alpha, \beta) \\ & - b f_{7,\beta}(\alpha, \beta)] f_{13}(\gamma). \quad (63) \end{aligned}$$

A good test of most interpolation schemes and wind field simulation techniques can be administered through the use of U and V in Eq. (54) and (55), with $f_2 = f_5 = 1$, and W in Eq. (63).

A candidate for f_2 and f_5 is the Ekman spiral given by

$$\begin{aligned} f_2(\gamma) &= 1 - \exp[-\pi\gamma] \cos[\pi\gamma], \\ f_5(\gamma) &= \exp[-\pi\gamma] \sin[\pi\gamma]. \end{aligned} \quad (64)$$

Candidates for the vertical shear term are

$$\begin{aligned} f_1(\gamma) &= \ln(\gamma + 1); \gamma^4 - \gamma^5; \gamma^{1/4} - \gamma^{1/3}; \\ &\gamma^{1/2} - \gamma; \gamma^{1/7}; 4/3\gamma^{1/4} - \gamma. \end{aligned} \quad (65)$$

The expressions f_4 and f_7 can be reasonable functions of α and β as long as Eq. (56) and (57) are satisfied. The functions f_3 and f_6 must satisfy Eq. (53), (56) and (57); that is f_3 must be the real part of an analytic function of $\alpha + i\beta$, while f_6 must be the negative of the imaginary part. Examples of f_3 and f_6 are given in Table 1.

Table 1. Examples of $f_3(\alpha, \beta)$ and $f_6(\alpha, \beta)$.

<u>$F(\alpha + i\beta)$</u>	<u>$f_3(\alpha, \beta)$</u>	<u>$f_6(\alpha, \beta)$</u>
$\alpha + i\beta$	α	$-\beta$
$(\alpha + i\beta)^2$	$\alpha^2 - \beta^2$	$-2\alpha\beta$
$(\alpha + i\beta)^3$	$\alpha^3 - 3\alpha\beta^2$	$-3\alpha^2\beta + \beta^3$
$\exp(\alpha + i\beta)$	$\exp(\alpha) \cos(\beta)$	$-\exp(\alpha) \sin(\beta)$
$\cos(\alpha + i\beta)$	$\cos(\alpha) \cosh(\beta)$	$\sin(\alpha) \sinh(\beta)$
$\ln(\alpha + i\beta)$	$1/2 \ln(\alpha^2 + \beta^2)$	$-\tan^{-1}(\beta/\alpha)$
-----	$2(\alpha^2 - \beta^2) + 3$	$-4\alpha\beta + 1$

The Domain of Computation

Once the transformed wind field is chosen, the actual wind can be specified if $\bar{z}_s$, L_x , L_y and $\bar{H}$ are known. The specific terrain fixes the $\bar{z}_s$, while L_x and L_y are chosen on the basis of computational convenience or horizontal resolution. The imaginary surface $\bar{H}$ may coincide with a measured inversion surface or may be arbitrarily specified. Some candidates for arbitrary $\bar{H}$'s are given below:

$$\bar{H} = (1 + a) \bar{z}_s; a = 1/10, 1/2, 1.$$

$$\bar{H} = a + \bar{z}_s; a = 10m, 50m, 100m, 1000m.$$

$$\bar{H} = a > (\bar{z}_s)_{\max}.$$

$$\bar{H} = \bar{z}_s + a \frac{(\bar{z}_s)_{\max}}{\bar{z}_s}; a = 10m, 50m, 100m.$$

$$\bar{H} = \bar{z}_s + a - b \sin [c (x^2 + \{y - \frac{1}{2}L_y\}^2)^{1/2}];$$

$$a > b > 0.$$

Mass-Conserving Interpolation Scheme

We propose a mass-conserving interpolation scheme based on the formulas given in Eq. (50) to (53). The formulas for U and V are given by

$$U = \sum_{n=0}^N a_n(\gamma) b_n(\alpha, \beta), \quad (66)$$

$$V = \sum_{n=0}^N c_n(\gamma) d_n(\alpha, \beta), \quad (67)$$

where $N + 1$ represents the number of wind stations or the locations of the measured wind profiles in and around the domain of computation. The basis functions b_n and d_n are polynomials in α and β ; the first 30 quantities are given in Table 2. Using the measured winds, the amplitude functions $a_n(\gamma)$ and $c_n(\gamma)$ can be determined at several vertical locations (values of γ). Thus, the "horizontal" components (U , V) will coincide with all "measured"

Table 2. Examples of the b_n and d_n Basis Functions

n	b_n	d_n	n	b_n	d_n
0	1	1	15	α^5	$\alpha^4\beta$
1	α	β	16	$\alpha^4\beta$	$\alpha^3\beta^2$
2	β	α	17	$\alpha^3\beta^2$	$\alpha^2\beta^3$
3	α^2	$\alpha\beta$	18	$\alpha^2\beta^3$	$\alpha\beta^4$
4	$\alpha\beta$	β^2	19	$\alpha\beta^4$	β^5
5	β^2	α^2	20	β^5	α^5
6	α^3	$\alpha^2\beta$	21	α^6	$\alpha^5\beta$
7	$\alpha^2\beta$	$\alpha\beta^2$	22	$\alpha^5\beta$	$\alpha^4\beta^2$
8	$\alpha\beta^2$	β^3	23	$\alpha^4\beta^2$	$\alpha^3\beta^3$
9	β^3	α^3	24	$\alpha^3\beta^3$	$\alpha^2\beta^4$
10	α^4	$\alpha^3\beta$	25	$\alpha^2\beta^4$	$\alpha\beta^5$
11	$\alpha^3\beta$	$\alpha^2\beta^2$	26	$\alpha\beta^5$	β^6
12	$\alpha^2\beta^2$	$\alpha\beta^3$	27	β^6	α^6
13	$\alpha\beta^3$	β^4	28	α^7	$\alpha^6\beta$
14	β^4	α^4	29	$\alpha^6\beta$	$\alpha^5\beta^2$

data points, will automatically conserve mass if there is no stratification, regional lifting or subsidence, and veering or backing of the wind, and will automatically give a vertical wind component $\bar{w}$ that is produced by the forcing of the terrain and the upper surface $\bar{H}$.

If the measured winds contain vertical velocity components due to stratification, regional lifting (subsidence), or veering (backing), then W is computed from Eq. (42). Thus, mass consistency is again guaranteed. The wind field $(\bar{u}, \bar{v}, \bar{w})$ over the original domain (Figure 1) is obtained from the application of the coordinate transformation to (U, V, W) .

Summary

This paper summarizes the concepts and interpretation of mass-consistent modeling for wind fields over complex terrain. The conclusion of this study is that interpolation of measured winds over an area are more important than making them mass-consistent. Hence, we propose an interpolation scheme that is automatically mass consistent if vertical winds are only produced by terrain-forcing. If this is not the case, mass consistency is produced by adjusting the "vertical" wind component in a transformed coordinate system. This modified modeling procedure will be tested against actual data in the near future.

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MASS-CONSISTENT WINDFIELDS - JULY 22 GEYSER's AREA

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ABSTRACT

The method for developing the windfields for the July 22, 1979 tracer experiment, conducted at the Geyser's area of California by Group G-8 of LASL, will be described. The terrain following mass-consistent windfield code called ATMOS1 was used in conjunction with a study of the physics of drainage flow using a boundary layer code called ATMOS3. The use of these windfields in a terrain following advection-diffusion calculation (ATMOS2) will be presented in a companion paper at this meeting.

I. THE TERRAIN FOLLOWING MASS-CONSISTENT CODE ATMOS1

The approach used in ATMOS1 is similar to that used by Sherman (1977) in the cartesian coordinate code MATTHEW. The windfield is made mass consistent by minimizing the squared variation of the windfield, subject to the constraint, the adjusted windfield is non-divergent (Sasaki, 1958). The principle difference in these two codes is that ATMOS1 uses terrain following or sigma coordinates, while MATTHEW relies on the usual cartesian coordinate system. The sigma coordinate is defined as

$$\sigma = \frac{Z_T - Z}{Z_T - Z_s} = \frac{Z_T - Z}{\pi} \quad (1)$$

where Z_T is the "constant" altitude at the top of the mesh and Z_s the terrain surface. It is possible to use a transformation

other than linear, and relax the constraint on a fixed upper boundary, but at present we have no plans to make these changes. In these transformed coordinates, the divergence-free condition $\nabla \cdot \bar{U} = 0$ becomes

$$\frac{\partial \pi u}{\partial x} + \frac{\partial \pi v}{\partial y} + \frac{\partial \pi \tilde{w}}{\partial \sigma} = 0 \quad (2)$$

where $\tilde{w}$ is the vertical velocity component in sigma coordinates of

$$\tilde{w} = \frac{1}{\pi} \{ \sigma [(Z_s)_x u + (Z_s)_y v] - w \} \quad (3)$$

This conservative form of the divergence (Eq. 2) is useful in forming the difference equations more accurately. At the terrain surface therefore

$$\tilde{w}_s = (Z_s)_x u + (Z_s)_y v \quad (4)$$

where $(Z_s)_x$ and $(Z_s)_y$ are the terrain slopes. (i.e., $(Z_s)_x = \partial Z / \partial x$, etc.)

The Sasaki variational integral is then

$$I = \int_V \left\{ \alpha_1^2 [(\pi u - \pi u_0)^2 + (\pi v - \pi v_0)^2] + \alpha_2^2 (\pi \tilde{w} - \pi \tilde{w}_0)^2 + \lambda \left[\frac{\partial \pi u}{\partial x} + \frac{\partial \pi v}{\partial y} + \frac{\partial \pi \tilde{w}}{\partial \sigma} \right] \right\} dx dy d\sigma \quad (5)$$

where λ is the Lagrange multiplier. The α_1 , α_2 are the Gauss precision moduli as defined by Sherman. We have not attempted a study of the effects of stability as included in the modifications of the α 's in a manner described by M. Dickerson (1975).

After writing the Euler-Lagrange equations to minimize I, one finds that many complex terms arise of the form

$$\partial^2 \lambda / \partial x^2, \partial^2 z_s / \partial x^2, (\partial z_s / \partial x)^2, \text{ etc.}$$

A staggered grid is then used to finite difference the Poisson equation in λ resulting in a minimum of truncation error. The over-relaxation techniques are then used to solve this equation and, usually, one obtains a convergence of less than 10^{-6} .

II. THE INITIAL WINDFIELD INTERPOLATIONS

At present we are using a fairly simple procedure for the interpolation of the observed winds in order to fill the computational mesh. There are two parts to the method: first, the vertical interpolation with altitude, and, second, the spreading out of the winds in x and y . If the data is from vertical sounders, like pibals or tethersondes, we linearly interpolate to the location of the coordinates. For the surface data we include a $1/R^2$ interpolation from the surrounding vertical sounders to the location of the surface measurement station. A possibly better method would be to interpolate up in altitude from the surface with the usual approximation: $u = u_0 (z/z_0)^n$, where n is some value depending on stability (usually near $1/7$), and u_0 is the measurement of the velocity at the height z_0 . After these winds are introduced into the conformal terrain mesh, we interpolate on $1/R^2$ for the grid positions. To

describe the windfields for the July 22 tracer experiment, we used some eight ground stations, one pibal and two tethersondes (Fig. 1). Since we have no information on the vertical component of the wind (w), we have set $\tilde{w}_0 = 0.0$ in the initial interpolated wind determinations (Eq. 5). The condition $\tilde{w}_0 = 0$ at $\sigma = 1$ is equivalent to $w = 0.0$, and insures that the mass flux cannot cross the terrain surface. We are now contemplating the use of pseudo drainage flow stations in order to develop better near-surface slope winds. These stations would be a combination of upper level flow fields, as observed by the tethersondes or pibals, and drainage flow theory contained in ATMOS3 that was discussed at the Atlanta meeting (January 1979). A test of these windfields will be the comparisons of the tracer experiments with and without added slope flow.

III. A STUDY OF "DRAINAGE FLOW" WINDS

Since it appears difficult to separate out from the observed windfield data, the "idealized" drainage flow due to the limited measurements, we propose to study drainage flow separately using our 1D model (ATMOS3). The model calculates the appropriate nocturnal cooling and subsequent heating from solar radiation given the correct latitude, longitude, and time of day. The transport of the long wavelength radiation is carried out taking into account the humidity, carbon dioxide, and ozone in the atmosphere. The calculations were started at 6 o'clock local time with an assumed surface temperature of 293°K . At present, we have not considered a method for

introducing combined synoptic and drainage flows into our 3D mass consistent windfield models.

Using the observed slope of $1/5$ for Anderson creek, we developed slope winds for local times of 10, 12, and 2 o'clock. These winds, including the transition to upslope daytime winds, are shown in Fig. 2. The upslope winds in the daytime are much stronger than the nocturnal winds that reach a maximum near 3-m/s at an elevation near 30 meters above the slope surface. If we drop the humidity to 10% from 30%, the maximum velocity increases to 4-m/s as the cooling increases.

We plan to adopt the drainage flow winds as our pseudo stations in ATMOS1, as shown in Fig. 3. The effects of these pseudo stations are still being studied.

IV. FINAL DISCUSSION OF JULY 22 WINDFIELDS

The windfields that we have developed for the July 22, 1979 tracer experiments carried out in the Geysers area of northern California by G-8 of the LASL are based principally on the observed winds, using a terrain following mass-consistent windfield code (ATMOS1). The data consisted of eight surface measurements as indicated by dots on the terrain diagram of Fig. 1, and the tethersondes and pibal stations as indicated by circles. There does appear to be a null in the windfield near the surface at Anderson Springs as evidenced in the velocity plots for level 10, the lowest level in terrain following coordinates (Fig. 4). This level is like three meters above the terrain in the valley and about one meter above the terrain near

the Socrates mine road. The vertical profiles located along Anderson Creek (Fig. 5), are shown in Figs. 6 to 8. Hopefully, the interpretation of these windfields in terms of the tracer experiments, including the Putah Creek releases, will help in understanding flows over the complex terrain in the Geysers area.

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Fig. 1. Locations of wind observations used in ATMOS1.
 0 tethersondes and pibal • surface measurements.

Fig. 2. Drainage flow velocities at selected times.
 Slope = 0.20, $z_0 = 1.0$, relative humidity
 = 30%.

Fig. 3. Drainage flow for possible pseudo stations in ATMOS1. Direction of slope 15° , S of E (Anderson Creek).

Fig. 4. Wind vectors at 12 o'clock, July 22, 1979, Geyser's area. K = 10, lowest σ level, height above Anderson Springs like 1 meter and 3 meters above surface at Socrates mine road. Note null winds around Anderson Springs.

Fig. 5. Location of vertical wind profiles from ATMOS1. Stations from top of Anderson ridge to near Anderson Springs sequentially.

Anderson Creek 7/22/79 24:00 Six Positions

Fig. 6. Wind speeds from ATMOS1. Locations in Fig. 6 are stations starting near Anderson Ridge (top-left) to a station near Anderson Springs (lower-right).

Anderson Creek 7/22/79 24:00 Six Positions

Fig. 7. Wind directions from ATMOS1. (Same format as Fig. 6).

Anderson Creek 7/22/79 24:00 Six Positions

Fig. 8. The vertical component of velocity (w) from ATMOS1 (result of $\nabla \cdot \mathbf{u} = 0$). (Same format as Fig. 6).

TRACER EXPERIMENT MODELING - JULY 22 GEYSER'S AREA

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ABSTRACT

The tracer release experiment conducted by Group G-8 of Los Alamos Scientific Laboratory (LASL) in the Geyser's area of northern California on July 22, 1979 is studied using a terrain following advection-diffusion code (ATMOS2) with winds interpolated from the observations (companion paper). Results indicate reasonable agreement with the time history of the sequential samplers located at S1 (Socrates Mine) and S2 (TVR) and the integrated collectors at 4 locations.

Model calculations have been carried out that attempt to simulate tracer release experiments at the Geysers Area. Comparisons with the experimental data provide a test of our current ability to model pollutant transport. The calculations are carried out with a windfield code (ATMOS1) and an advection-diffusion code (ATMOS2). The resulting concentrations permit prediction of tracer particle collections at the measurement stations. Comparison of the observations and predictions of sequential and integrated counts shows generally good agreement, but with discrepancies in some instances. The method appears to be a useful one for pollutant transport predictions and for parametric studies (such as wind station location).

MEASUREMENTS

The tracer experiments by LASL Group G-8 in Anderson Creek Valley, California have been described in detail elsewhere (Clements et al. 1979, 1980). In summary, the experiments consisted of fluorescent powder releases (500 grams over approximately 20 minutes) and collection on type H Rotorods. We have utilized the measurements made on July 22-23 and July 24-25, 1979. Comparison is made with integrated collections (several hours) at four measurement locations and sequential collections (10 minutes each) at two locations.

WINDFIELDS

The windfields used for these calculations were based upon the results of fitting to the observed wind measurements by ATMOS1. The procedure is described by Davis and Bunker (1980). The code develops a mass-consistent windfield from the observations. Cells are 190 m x 190 m in the horizontal plane (25x31 cells) and with a σ -coordinate scheme in the vertical direction having cells ranging from about 1 meter in height at the surface to 300 m at the upper boundary (10 cells). We have assumed fixed winds during the periods 22:00 to 24:00 and from 24:00 to 02:00 (P.S.T. throughout).

TRANSPORT CALCULATIONS

The transport modeling is performed by the code ATMOS2 that carries out advection and diffusion calculations utilizing the

ATMOS1 windfields. Advection and diffusion transport are solved through a standard continuity equation in finite difference form. Some characteristics of the ATMOS2 calculations are:

1. a σ -coordinate scheme in the vertical direction with an implicit solution to avoid time steps becoming too short;
2. a Crowley 2nd order procedure to limit numerical diffusion;
3. boundary conditions include a set of buffer cells on the lateral and upper surfaces into which diffusion does not occur and through which the pollutant cannot pass;
4. diffusivities are based upon the algorithms of Smith and Howard (1972) and the work of Lantz et al. (1971);
5. Pasquill stability Class E is used throughout;
6. no precipitation or surface trapping of particles in the present form of the code; and
7. cell structure identical to that employed in ATMOS1.

A simple test indicates conservation of particles to within a few hundredth percent over a 3-hour period. Rotorod collector predictions are carried by appropriate integrations over volume and time and assuming no wind effect on collection rate at the Rotorod. A value of 2.31×10^{10} particles per gram of dispersed material is assumed. The calculations with ATMOS2 require about 40 seconds of 7600 time per hour of model time.

CINE GRAPHICS

A computer-generated movie has been prepared of the tracer transfer calculations for the July 22-23 experiment. Time steps range from 23 to 28 seconds and the duration covers the pertinent portion of the 4-hour experiment. Each particle of the plots represents 0.1 grams of tracer. The movie permits an appreciation of the predicted flow characteristics including such features as the general flow down Anderson Creek and the formation of the tracer plume.

RESULTS

Figure 1 indicates the area included in these calculations, the release point, and the measurement stations. Figures 2 and 3 show the sequential measurements and predictions at stations S1 and S2. Both observations and predictions are normalized to unity at maximum. The time interval is 10 minutes for each collection with zero time at first release. Both the shapes and timing of the calculated curves show reasonably good agreement with measurements. The S2 July 24-25 measures show a conspicuous late-time plateau that is not predicted.

Figure 4 shows the integrated results for all stations. Experimental or calculated counts at each station have been divided by those at station I1 in each instance, thus providing an indication of count falloff from the station nearest the release point. It is seen that the calculations for July 22-23

show excellent agreement with the observations. By contrast, the comparisons for July 24-25 are very poor. At the present time there is no clear understanding of the causes of these differences. In a preliminary effort to understand the sensitivity of the predictions to the windfield, we have carried out a windfield calculation for July 24-25 in which the tether sonde data has been eliminated and two downslope wind pseudo stations have been added (based upon results from our 1D code ATMOS3). Figure 5 shows the resulting predictions, together with the original predictions, and it is seen that the comparison with observations is greatly improved.

CONCLUSIONS

It appears that good predictions of relative pollutant levels are possible at considerable distances from the source. Clearly, a careful analysis of the windfields and measurements will be necessary to explain discrepancies encountered in the integrated counts for July 24-25. As an example, there is an indication of a substantial difference in the upper-level wind patterns between the two measurement periods. Our trial adjustments of the windfields show a strong requirement for a careful use of the wind measurements.

Future improvements in the calculations should include an improvement of the diffusion and stability calculations and an attempt to model particle precipitation. ATMOS2 should be a useful tool for a better understanding of the effects of windfield fitting and of assumed parameters on pollution transport modeling.

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RELEASE POINT: ⓧ R1 ANDERSON CREEK

OBSERVATION POINTS:

- △ I1, S1 SOCRATES MINE ROAD
- I2 ANDERSON SPRINGS
 (1.45 MILES FROM HW 175)
- ◇ I3 ANDERSON SPRINGS
 (0.7 MILES FROM HW 175)
- ▽ I9, S2 TV REPEATER SITE

P-7

LOS ALAMOS SCIENTIFIC LABORATORY

ANDERSON CREEK DRAINAGE

Figure 1.

July 22, 1979 22:00 Release FP2

Time (min)

Figure 2.

July 24, 1979 22:25 Release FP2

Figure 3.

Integrated Sampling Stations

Figure 4.

Integrated Sampling Stations

**MODELING A TRACER RELEASE EXPERIMENT DURING
NOCTURNAL DRAINAGE FLOW IN COMPLEX TERRAIN**

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ABSTRACT

The three-dimensional particle-in-cell advection and diffusion code, ADPIC, was used to model two simultaneous, distinguishable fluorescent particle releases into the nighttime drainage flow regimes of the Putah, and Anderson Creeks at the Geyser's geothermal site in California during the night of July 22, 1979.

Two approaches were taken. First, the three-dimensional advection field (mean winds) which ADPIC needs as input was provided by the diagnostic mass conservative wind field model MATHEW. Second, the advection field was provided by a dynamic second order closure boundary layer model (Yamada) which also provided the vertical diffusivity parameters. The modeling and experimental results were compared.

INTRODUCTION

In order to better understand the atmospheric effects of energy production on the local and regional scale, the U. S. Department of Energy (DOE) has established the program of Atmospheric Studies in Complex Terrain (ASCOT). One aspect of this effort is to improve the numerical modeling capabilities for the study of transport and diffusion of aerial pollutants over terrain typical of energy development sites. To provide the data, an initial two week ASCOT field study was conducted in July 1979 in the Geysers geothermal area of California 100 km north of San Francisco to study nocturnal drainage flow in a series of six nighttime experiments.

The complex conditions posed by very localized winds and turbulence, rugged topography, variable surface roughness, strong vertical motions and gravity flow require a three-dimensional modeling approach to realistically simulate the dispersal of hydrogen sulfide injected into the atmosphere at the Geysers geothermal area. In this study the three-dimensional particle-in-cell numerical diffusion model ADPIC (Lange, 1973 and 1978) was used to simulate the transport and diffusion of two simultaneous, distinguishable fluorescent tracer releases into the nighttime drainage flow regimes of the Putah and Anderson Creeks in the Anderson Springs Valley during the night of July 22, 1979.

The ADPIC code requires a three-dimensional mean wind field as input and there are currently two significantly different classes of models capable of supplying such a wind field: the mass-consistent diagnostic models, and the first or second order closure dynamical models. In general comparison, present mass-consistent models have the advantage that they can use data directly, not only on the boundaries of, but also within their domain, are easy to update and use less computer time; but they incorporate not much physics and are prognostic only in the

sense of persistence. Present dynamical models contain the physics to be both diagnostic and prognostic. Second order types can provide local diffusivities, and they allow better understanding of the separate physical processes; but they have limited flexibility in their initial and boundary conditions and in the use of data internal to their domain, and they are costly on the computers.

In an effort to compare the wind fields produced by the two different approaches, ADPIC was run twice for the same tracer scenario: once using the wind field produced by the mass conservative MATHEW (Sherman, 1978) model and second using the windfield produced by the second order dynamical model of Yamada (Yamada, 1980). The results were then compared in regard to the differences in the wind fields and in a qualitative sense with the results of the tracer experiment.

Of the input models to ADPIC, the MATHEW model and its output wind field will be discussed in some detail. For a discussion of the Yamada model and its output wind field the parallel ASCOT report (Yamada, 1980) should be consulted.

THE JULY 22, 1979 EXPERIMENT

The experiment consisted of obtaining meteorological and fluorescent tracer release data over the approximately 10 km by 7 km area of the bowl-shaped Anderson Springs Valley during the nighttime drainage flow regime. The experiment began at 2200 PST and ended at approximately 0500 PST on the 23rd, although, some data were collected on a 24 hour basis. The Anderson Springs Valley is surrounded by Boggs Mountain on the northeast, by Cobb Mountain on the north, and by Pine Mountain on the southwest, and forms a bowl shaped drainage regime for the Putah, Gunning, Anderson and Bear Creeks opening towards the east-southeast. It has rugged topography and a ground cover varying from bare soil to forest canopy. Figure 1 shows the geography of the Anderson Springs Valley, the

measurement, tracer release, and sampler locations, and the outline of the ADPIC grid. The type of data connected with the symbols on the map are listed in Table 1, which shows the measurements used for the MATHEW/ADPIC codes. (Measurements not directly usable by these codes such as humidity and radiation fluxes are not shown but can be found in the 1979 ASCOT Experiment report.) Table 1 also shows the data for the two fluorescent particle releases. Release height was approximately 1.5 m.

CODE DESCRIPTION

MATHEW (Sherman, 1978) provides a three-dimensional gridded mean wind field, which has been built up from regionally available interpolated meteorological data and then has been made mass conservative in the least square sense subject to the constraints of the continuity equation and flow boundary conditions. The boundary conditions consist of zero flow through the topographical surface and full in and outflow through all other surfaces of the grid. Horizontal versus vertical mass fluxes needed to provide mass conservation for each cell can be adjusted by a flow ratio coefficient in order to take unstable/stable atmospheric conditions into account. The input for the model consists of a digitized topographical surface, spatially interpolated surface winds, and vertical wind profiles.

The Yamada (Yamada, 1978 and 1979) model, a numerical three-dimensional second-moment closure model, is the second model used to provide a mean wind field for the ADPIC transport and diffusion calculations. In addition, it provided the vertical diffusion coefficients for ADPIC through its second-moment parameterization of the mass fluxes. Details of the model and results for the Geysers experiment are reported elsewhere in this ASCOT Report (Yamada 1980).

ADPIC (Lange, 1973 and 1978) is a three-dimensional numerical particle-in-cell diffusion and transport model capable of simulating the time and space varying dispersal of atmospheric pollutants in the presence of complex meteorological and topographical conditions. The method is based on the particle-in-cell technique by which Lagrangian "mass" particles are transported inside a fixed gridded Eulerian velocity grid. In essence, the model solves the three-dimensional diffusion-advection equation in flux conservative form (pseudo velocity technique) for a given non-divergent advection as mean wind field:

$$\frac{\partial \chi}{\partial t} + \vec{\nabla} \cdot (\chi \vec{U}_P) = 0 \quad (1)$$

Here χ is the pollutant concentration and $\vec{U}_P$ a pseudo velocity which can be defined as being made up of the given mean advection or wind field $\vec{U}_A$ and a diffusive velocity $\vec{U}_D$ through the relationship

$$\vec{U}_P = \vec{U}_A + \vec{U}_D \quad (2a)$$

where

$$\vec{U}_D = -K \frac{\vec{\nabla} \chi}{\chi} \quad (2b)$$

and K is the diffusivity in the x , y and z directions. It is the $\vec{U}_A$ advection term (mean wind field) in mass conservation form that is supplied to ADPIC by other models.

The Lagrangian "mass" marker particles are transported through the grid with the Eulerian grid point pseudo velocities $\vec{U}_P$, interpolated to each particle location within a grid cell. It can be seen then from Eqs. (2a) and (2b) that the particle motion depends on a mean velocity $\vec{U}_A$, the local particle concentration χ and gradient $\vec{\nabla} \chi$, and the diffusivities K . ADPIC computes a horizontal and a

vertical diffusivity K_h and K_z in a variety of forms based on accepted diffusion theory. In this study the horizontal diffusivity K_h is based on the semi-empirical expression (Draxler, 1976)

$$\sigma_y = \sigma_\theta \bar{U} t f\left(\frac{t}{\tau}\right); \quad f\left(\frac{t}{\tau}\right) = \left(1 + \left(\frac{t}{\tau}\right)^{1/2}\right)^{-1} \quad (3a)$$

combined with the analytical relationship

$$K_h = \sigma_y \frac{d\sigma_y}{dt} \quad (3b)$$

where σ_y is the horizontal standard deviation of the pollutant plume width, σ_θ the standard deviation of the fluctuation of the wind direction, $\bar{U}$ the mean wind speed, t is time, and $f(t/\tau)$ is a function of time and an empirical time constant τ which will assure that Eq. (3a) will go to the Taylor limit $\sigma_y \propto t^{1/2}$ for times $t \gg \tau$. Note that K_h is a function of height also, through the height dependence of σ_θ .

The vertical diffusivity K_z is computed in ADPIC separately for the atmospheric surface layer and the outer boundary layer. For the surface layer K_z is based on similarity theory (Businger et al., 1971)

$$K_z = \frac{kU_*z}{\phi(z/L)} \quad ; \quad (4)$$

where k is the Von Karman constant, U_* is the friction velocity computed by the code, z is the height above terrain, and $\phi(z/L)$ is an atmospheric stability function based on z and the Monin-Obukhov scale length L . In the outer atmospheric boundary layer K_z is computed by the empirical form

$$K_z = CU_*z e^{-(z/h)^4} \quad (5)$$

where C is a proportionality coefficient obtained from data fitting (Start et al., 1975; Crawford, 1974), and h is the mixing height of the atmospheric boundary layer. With the exponent $-(z/h)^4$ maximum K_z occurs approximately at a height of $z/h = 0.8$.

CODE INPUTS

MATHEW Input

The grid consisted of $48 \Delta x \cdot 37 \Delta y \cdot 23 \Delta z$ cells with $\Delta x = \Delta y = 254$ m and $\Delta z = 50$ m and the long side of the grid in the east-west direction. Figure 2 shows the computer generated model topography for MATHEW and ADPIC. ADPIC uses a smaller grid than MATHEW in order to minimize boundary effects.

The surface wind measurements, as listed in Table 1 and whose locations are shown in Fig. 1, were all used as input with the exception of 101, W5 and M. The three mentioned had to be left out because at present MATHEW input is limited to 25 surface stations. The station winds were horizontally interpolated to grid points at a reference height of 5 m above topography. They were reduced to this height via a 0.30 exponent power law. A surface roughness height of 0.2 m was assumed. Figure 3 shows the station wind vectors at the reference height.

This interpolated horizontal grid was then tied to a vertical profile which was a visual compromise of the four vertical profiles measured by the pibal, two tether sondes, and the WPL double acoustic sounder as shown in Fig. 4. The figure shows that one vertical alone profile is not a good approximation of the data, however, at present MATHEW is limited to using one vertical profile although it can have up to five different wind inputs at the top of the grid; both those and the surface winds are tied to the vertical profile by interpolation. Three hourly averaged data sets were prepared for MATHEW to produce mass conservative wind

fields for 2200, 2300 and 0000 PST. Where the data were given for less than hourly averages, such averages were prepared.

ADPIC Input

The grid has the same cell size and topography as MATHEW, however, it is smaller in the horizontal $38 \Delta x \cdot 27 \Delta y$ but of the same height. Because ADPIC accepts the three-dimensional gridded wind field from MATHEW as input, the only major parameters that are still needed are diffusion, and pollutant source data. Figure 5 shows the σ_θ profile compiled from data near the surface and from Turner type stability considerations aloft; this is the σ_θ used in ADPIC in the horizontal dispersion calculation based on Eq. (3a).

For the vertical diffusivity based on Eqs. (4) and (5), the M.O. length L was chosen as 25 m and the mixing height h was chosen as 500 m based on the temperature profiles in Fig. 4. The stability coefficient C in Eq. (5) was obtained by normalizing the maximum K_z from this equation to those obtained by applying in the vertical, the generally valid analytic relationship between K and σ , Eq. (3b) and using σ_z data of complex terrain modified Pasquill curves (Start et al., 1975).

The sources were modeled in ADPIC by sequentially releasing 10,000 simulative tracer particles each, at the two locations indicated as R1 and R2 in Fig. 1 over the release period indicated in Table 1.

Yamada Model Input

Essentially the same data were used as input as were used for MATHEW (Yamada 1980, ASCOT Report). The Yamada model grid was slightly different from that chosen for MATHEW, both in size and number of cells: $27 \Delta x \cdot 20 \Delta y \cdot 24 \Delta z$, $\Delta x = \Delta y = 381\text{m}$, $\Delta z = 50\text{m}$. The terrain following and vertically variably gridded wind field produced by the Yamada model was interpolated to the Cartesian grid

block needed as input by ADPIC. The ADPIC grid was chosen to conform with the Yamada grid but all the other ADPIC inputs remained the same as those used with MATHEW winds with the one exception that the vertical diffusivities K_z were provided in a gridded array by the second moment closure parameterization of the Yamada model.

RESULTS

Both the MATHEW/ADPIC and the Yamada/ADPIC computer runs were over in little more than an hour (real-time) in the sense that all released particles had been transported out of the grid in that time span. This means that little more than the first 2200 PST hourly wind field contributed to the flow, thus making the advection part of the transport essentially a steady state flow. Furthermore, the 2300 PST wind sets did not differ much from the previous; so only the 2200 PST MATHEW wind field set will be discussed. Then will follow the results of the MATHEW/ADPIC and Yamada/ADPIC runs. The Yamada wind field is presented other places in this report (Yamada, 1980).

The MATHEW mass-adjusted wind field for 2200 PST in wind vector form is illustrated in Figs. 6, 7, and 8. For clarity, only every third grid point is plotted. The surface winds are light, and even calm over the center of the valley, but are fairly strong near the Anderson and Putah Creek tracer release points. This is the result of the two Livermore 10 m wind towers L6 and L7, see Fig. 3, and is probably the reason for the ADPIC particles leaving the grid in little more than one hour's time. The latter is true both for the MATHEW/ADPIC and the Yamada/ADPIC runs, although the sequential surface tracer measurements at S1, S2, and S3 in Fig. 1 indicate a slower transport of the fluorescent particle releases, a discrepancy which will be discussed later. The dashed line in Fig. 7 indicates the location of a vertical

cross section of the adjusted velocity field grid, through the Anderson Valley in the east-west direction. This section is shown in Fig. 8 and gives an indication of the vertical airflow within the valley as provided by MATHEW. (The plot appears to violate mass conservation, but it is to be remembered that it shows only two of three dimensions.) The vertical divergence of the air flow is caused by a strong horizontal convergence, as the winds flow downhill into the bowl-shaped valley from over the westerly and southerly ridges and through the Putah Creek gap between Cobb and Boggs mountain, and then are restricted in their outflow through the southwesterly drainage opening of the valley. The model produced maximum upward velocities on the order of 1 m/s at the location. The Yamada wind field used shows similar features.

A better appreciation of the horizontal convergence of the winds can be gained from Figs. 9, 10, and 11 which show the flow lines of the adjusted MATHEW wind field for 2200 PST for various heights above the topography. There is an indication of drainage flow from the southwesterly rim of the valley and through the Putah Creek gap. The maximum in the vertical wind profile due to this flow occurs at about 200 m above terrain as can also be seen in Fig. 3. Of all the profiles measured, only the LASL profile at the TVR indicates a low level maximum, all the others, including the compromise profile used in the calculation, have maxima at higher elevations which cannot clearly be identified as drainage flow. There may be a drainage flow present that has "lifted off" by sliding over yet colder air already settled in the valley due to the complexity of the topography, which makes the Anderson Creek Valley a region of multiple drainage flow regimes. Such layering of the flow is indeed observed and seems to be supported by the computer runs with ADPIC by the paths followed by the simulated tracer releases. The Yamada wind field shows profile maxima at lower heights, more characteristic of drainage flow.

The results of the MATHEW/ADPIC calculation are shown in Figs. 12 to 18 which represent the transport and diffusion of the simulated Anderson Creek and Putah Creek releases through the ADPIC grid at twenty minute intervals. Figures 12, 14, 16, and 18 are views "downward" through both plumes of ADPIC particles. Figures 13, 15, and 17 represent views "to the north" through the valley and the entire particle plumes with the topography under the plume centerlines indicated. For better model intercomparison, the release plume centerline trajectories are indicated, the precursors being caused by the difference between problem start at 2200 PST and start of release at 2215 PST. A similar sequence of Figs. 19 to 23 shows the corresponding Yamada/ADPIC results.

Comparing the advective aspects of the flow, i.e., the wind fields, via the centerlines of the modeled ADPIC plumes there is an overall agreement between the MATHEW/ADPIC and Yamada/ADPIC in the major features of these trajectories. While details differ, the Yamada/ADPIC Anderson and Putah Creek plumes cross paths at different elevations (Figs. 21 and 22), the plumes in both ADPIC runs flow out of the valley within 1 km of Diamond D Ranch. Both runs indicate the Anderson Creek plumes are "lifting off" the surface as they approach the center of the bowl-shaped valley. As mentioned earlier, it may be that this is caused by the strong horizontal convergence of the airflow as colder air descends into the valley from its perimeter, and the ensuing vertical upflow over the center of the valley due to the restricted outflow along Highway 175.

A qualitative comparison of the plume diffusion between the two ADPIC runs can also be gained from the sequence of Figs. 12-23. The prescription for horizontal diffusion coefficients discussed in the description of the ADPIC model was used for both MATHEW/ADPIC and Yamada/ADPIC. The chief difference evident between the horizontal plume spread results from the stronger horizontal convergence and higher speeds of the Yamada wind field. Especially, the narrowing of the horizontal

extent of the plumes combined with increased vertical spread at the same location in the Yamada/ADPIC in Figs. 21 and 22 is an illustration of this effect, which can occur in ADPIC because of Eq. (2). Vertical diffusion coefficients were not the same: the MATHEW/ADPIC used Eqs. (4) and (5), while the Yamada/ADPIC used K_z 's computed from the second order closure condition. It appears that the ADPIC K_z 's were in general larger than those provided by the Yamada model, but neither model had strong enough vertical diffusion to overcome the detachment of the model plumes from the surface so that the model surface samplers did not "see" the plume.

The comparison of MATHEW/ADPIC and of Yamada/ADPIC with the experimental results of the fluorescent particle (FP) releases into the Anderson and Putah Creeks can be qualitative only, largely because the behavior of FP as a neutrally buoyant mass balance tracer is not established: field samplers were surface samplers only and some of the larger samples only contained on the order of 10^3 particles out of a total of 10^{13} released. Whether the bulk of the FP "lifted off" as the model runs would indicate, or got attached to the canopy is not known, although the latter is more likely due to their slightly negative buoyancy.

The model plume particles for both ADPIC runs passed the sequential sampler locations S1, S2, S3 (Fig. 1) within one hour of release while the real FP took three hours, see Fig. 24. This is probably due to the fact that the canopy drag effect included in the models was not sufficiently large enough. Figure 24 shows time of overhead passage of the MATHEW/ADPIC and Yamada/ADPIC Anderson and Putah Creek releases at S1, S2, and S3 together with the experimental results. Figure 25 is an overlay of the MATHEW/ADPIC plume centerlines at the end of the code runs over a quasi-quantitative description of the total integrated field samples taken of the Anderson and Putah Creek FP. The size of the dots is roughly proportional to the sample size. I6, I7, and I8 are integrated samplers which were pumping until

0400 PST. These samplers lie in Putah Creek and how Anderson Creek particles got there is not explained by the model runs. However, had the model runs been longer, i.e., had the particles been delayed by canopy effect, they might have found their way into Putah Creek, because later MATHEW hourly wind fields indicated some back flow into Putah Creek.

CONCLUSIONS

From the results of this study, conclusions can be drawn along two major lines: model intercomparison and model comparison with experimental results of the tracer experiment.

The fact, that two different three-dimensional models like the mass conservative, diagnostic MATHEW model and the second-order dynamical closure model of Yamada agreed in the main features of the air flow regime in the complex terrain of the Anderson Creek Valley during a nighttime drainage scenario, is encouraging. They both featured the strong horizontal convergence of cold air descending into the bowl-shaped valley from its perimeter and the ensuing vertical updraft over the center and the terrain limited outflow region of the valley along Highway 175. A result was detachment from the surface of the Anderson Creek ADPIC plume as it approached the valley center most likely due to colder air already present in the valley from other drainage features of this complex drainage regime. Such layering of air masses of different temperatures was observed in the field, especially by the acoustic sounder at the Diamond D. Ranch. Unfortunately, proof, if any, of such possible plume "lift off" awaits the 1980 Geysers tracer experiment.

While both models show the presence of drainage flow, the MATHEW model has its maximum in the vertical wind profile at about 150 to 200 m above terrain and its

identification as gravity flow may be disputable. Perhaps a smaller vertical grid spacing than the 50 m used is needed to be able to resolve a shallow drainage flow. On the other hand, MATHEW-type models can only produce drainage flow profiles from conservation of mass, measurements, and topography, and the night of July 22 was a weak drainage night and only one of the four measured vertical wind profiles (Fig. 4) displayed a maximum at 50 m height. The average profile used for MATHEW did not have this maximum which indicates that for future drainage wind studies in complex terrain, it is necessary to be able to handle multiple profiles in MATHEW.

Because of the questionable behavior of fluorescent particles as a mass balance tracer in this case, and the limited sampling, satisfactory conclusions with respect to the diffusion parameters in ADPIC and the Yamada models cannot be drawn. In a gross sense, if the advective trajectories of the MATHEW and Yamada model plumes including the "lift off" are to be believed, then the diffusivities in both models were too small, because, while the field surface samplers actually did observe FP from both the Anderson and Putah Creek releases, the ADPIC run for neither model showed surface concentrations at the sampler location.

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TABLE 1. Data used to model July 22, 1979 episode.

Symbol	Organization	Instrument	Measurements*
<u>Meteorological Data</u>			
TVR	LASL	Pibal, acoustic sounder	S, θ, T
101-105	LASL	Surface stations	$S, \theta, T, \sigma_{\theta}$
W1-W12	LASL	Van stops	$S, \theta, T, \sigma_{\theta}$
T1, T2	PNL/WPL	Tethersondes (~550 m)	S, T, θ
L1-L7	LLNL	Towers (10 m)	$S, T, \theta, \sigma_{\theta}$
16, 19	PG&E	Towers (10 m)	S, θ, T
P1	PNL	Tower (40 m)	$S, T, \theta, \sigma_{\theta}$
OL	LLNL	Laser	$S, \theta, (\sigma_{\theta})$
M	MRI	Surface station	S, θ, T
DD	WPL	Acoustic sounder	S, θ
<u>Release Data</u>			
R1, R2	LASL	Fluorescent particle release location	
I1-I5	LASL	Fixed samplers (integrated)	
S1-S3	LASL	Spot samplers	
PN	PNL	Integration and sequential sampling array	

Start of Releases: 2215 PST.

Anderson Creek Release: 500 g yellow. 1.16×10^{13} particles over 23 min.

Putah Creek Release: 500 g green. 1.16×10^{13} particles over 25 min.

* S = Speed

θ = Direction

T = Temperature

σ_{θ} = Standard deviation of horizontal wind fluctuation

FIGURE 1. Geographical map of the Geysers geothermal area Anderson Springs Valley and meteorological instrument and sampler locations of the July 1979 ASCOT experiments.

Located
at TVR
101
L1
S2
I9

Located
at TH7
T1
L3

Located
at ACR
102
W3
I1
S1
PN

FIGURE 2. Computer generated topography of Anderson Springs Valley for the MATHEW/ADPIC calculation. The contour lines are in meters above the lowest point of the grid.

FIGURE 3. Wind velocity data for 7/22/79, 2200 PST adjusted to a height of 5 m above topography using power law exponent of 0.30.

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FIGURE 4. Measured and model input vertical wind profiles above topography for 7/22/79, 2200 PST. See Table 1 and Fig. 1 for identification and location.

FIGURE 5. Profile of the horizontal standard deviation σ_θ in degrees used in the ADPIC dispersion calculation. Dots are input, solid line is curve fit.

FIGURE 6. Adjusted wind velocities at the surface. Vectors are plotted at every third grid point.

FIGURE 7. Adjusted wind velocities 450 m above topography. Vectors are plotted every third grid point. The dashed line indicates the location of the vertical cross section through the grid shown in Fig. 8.

FIGURE 8. Adjusted U-W wind velocities at a vertical section through the grid as located by the dashed line in Fig. 7. Scale of W components = 100 * scale of U components.

FIGURE 9. Flow lines of the adjusted surface winds. Density of arrows is proportional to wind speed.

FIGURE 10. Flow lines of the adjusted winds 50 m above the surface. Density of arrows in proportional to wind speed.

FIGURE 11. Flow lines of the adjusted winds 450 m above the surface. Density of arrows is proportional to wind speed.

FIGURE 12. MATHEW/ADPIC plumes at time = 2220 PST. Trajectories passing through release points are shown in addition to actual releases. A = Anderson Creek release, P = Putah Creek release. S1-S3 are sequential samplers.

FIGURE 13. MATHEW/ADPIC plumes in side view looking north. Time = 2220 PST. Trajectories passing through release points are shown in addition to actual releases. Topography under the plume centerline is indicated for Anderson and Putah Creek plumes.

FIGURE 14. Same as Fig. 12 but at time 2240 PST.

FIGURE 15. Same as Fig. 13 but at time = 2240 PST.

FIGURE 16. Same as Fig. 12 but at time = 2300 PST.

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FIGURE 17. Same as Fig. 13 but at time = 2300 PST.

FIGURE 18. Same as Fig. 12 but at time = 2320 PST.

FIGURE 19. Yamada/ADPIC plumes at time = 2220 PST. Plume centerline trajectories are shown in addition to actual releases. A = Anderson Creek release. P = Putah Creek release. S1-S2 are sequential samplers.

FIGURE 20. Yamada/ADPIC plumes in side view looking north. Time = 2220 PST. Plume centerline trajectories are shown in addition to actual releases. Topography under plume trajectory is indicated for Anderson and Putah Creek plumes.

FIGURE 21. Same as Fig. 19 but at time = 2240 PST.

FIGURE 22. Same as Fig. 20 but at time = 2240 PST.

FIGURE 23. Same as Fig. 19 but at time = 2300 PST.

FIGURE 24. Experimental plume passage times at sequential samplers S1, S2, and S3 compared with MATHEW/ADPIC and Yamada/ADPIC plume passage intervals for comparison. (For the code results, time of passage above sampler location.) Solid line = Anderson plume, dashed line = Putah plume.

FIGURE 25. MATHEW/ADPIC and Yamada/ADPIC plume centerline trajectories for Anderson and Putah Creek plumes at time = 2300 PST, and approximate size of integrated field samples taken. Size of dots is roughly proportional to total number of fluorescent particles trapped.

A STATISTICAL MODEL FOR WIND PREDICTION AT A MOUNTAIN AND VALLEY STATION NEAR ANDERSON CREEK, CALIFORNIA

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1. INTRODUCTION

Studies of relations between the geostrophic wind, heat transfer and surface wind are basic in the understanding of the mechanism for the maintenance of the kinetic and thermal energies associated with the mean and turbulent motions in the atmospheric boundary layer. For example, Lettau (1959), Kazanski and Monin (1960), Csanady (1967), Blackadar and Tennekes (1968), and Brown (1974) have analyzed the relations for the atmospheric and oceanic boundary layers under stationary, horizontally homogeneous and neutral stability conditions. Zilitinkevich, Kaikhtman and Monin (1967), Zilitinkevich and Chalikov (1968), Clarke (1970, 1972), and Arya (1975) have extended the investigation to include the effect of vertical heat flux. However, all these studies have emphasized the geostrophic drag relations in the atmospheric boundary layer over flat surfaces.

The attempt of this paper was to seek statistical relations between the geostrophic wind, vertical heat flux and surface wind, under nonstationary conditions, in the atmospheric boundary layer near Anderson Creek, California (Fig. 1).

2. WIND SPEED MODEL

Winds in complex terrain are greatly affected by the topography, pressure gradient force, thermal stability and cross isobaric angle at the surface. To formulate a statistical equation for the surface wind speed in complex terrain, we assume that the ratio of the surface wind speed $V(t)$ to the geostrophic wind speed $G(t)$ takes the following form:

$$\frac{V(t)}{G(t)} = a_1 \frac{V_s(t)}{G(t)} + a_2 \frac{V_v(t)}{G(t)} + a_3 \alpha_o(t) + a_4 \mu(t) + a_5(t) \quad (1)$$

where $V_s(t)$ is the slope wind speed, $V_v(t)$ is the valley wind speed, $\alpha_o(t)$ is the cross isobaric angle, $\mu(t)$ is a nondimensional thermal stability parameter defined as

$$\mu = \frac{ku_*}{|f|L} \quad (2)$$

where k is the Von Karman constant, u_* is the friction velocity, $f = 2\Omega \sin \phi$ is the Coriolis parameter, and

$$L = - \frac{c_p T \rho u_*^3}{kgH} \quad (3)$$

is the Monin-Obukov's length, in which c_p is the specific heat at constant pressure, T is the temperature, g is the gravity acceleration, ρ is the air density, and H is the vertical heat flux.

The valley wind speed is approximated by the following Fourier series:

$$V_v(t) = \bar{V}_v + \sum_{i=1}^3 \left\{ A_i \cos \left(\frac{2\pi}{24} it \right) + B_i \sin \left(\frac{2\pi}{24} it \right) \right\} \quad (4)$$

where $\bar{V}_v$ is the mean valley wind speed, A_i and B_i are amplitudes of the i th harmonic of the Fourier series. In general, $\bar{V}_v$, A_i and B_i are functions of location and season; their values for mountain station SRI1 and valley station SRI5 are listed in Table 1.

The slope wind is generally a function of the slope of the terrain, time of day, solar insolation and sky condition. The solution is based primarily on relationships developed by Petkovsek and Hocevar (1971) and Ryan (1974). Thus

$$V_s(t) = 0.895 \frac{(I - C_n)}{\tan S} \left\{ 1 - \exp(-1.5 \Delta t \tan^2 S) \right\} \quad (5)$$

where I is the solar insolation, C_n is a function of nighttime sky covering with $C_n = 0.67$ for cloud cover less than 0.5 and $C_n = 0.34$ for cloud cover greater than 0.5, Δt is the elapsed time in decimal hours since sunrise for upslope flow or since two hours before sunset until two hours after sunrise for downslope flow, and S is the slope of the terrain.

The values of coefficients a_1 , a_2 , a_3 , a_4 , and a_5 for the mountain and valley stations for the eight time periods (three hours for each period) since midnight are listed in Tables 2 and 3. It may be noticed that the surface wind speeds at the mountain and valley stations are primarily affected by the valley, slope, and geostrophic winds. Meteorological data collected near Anderson Creek, California, from August 1 to August 17, 1977, were used for this study.

The observed surface wind speed at the mountain and valley stations for the period from August 13 to August 17, 1977, are shown as solid curves in Figs. 1 and 2. The surface wind speeds calculated with the use of Eq. (1) are shown as dashed curves in Figs. 2 and 3. It may be noticed that calculated wind speeds generally agree well with the observed; their correlations for the eight time periods are listed in Table 4.

3. WIND DIRECTION MODEL

Because the variability of wind direction, which is greatly affected by topography, geostrophic wind and turbulence intensity, are generally larger than that of wind speed, statistical models for wind direction prediction in complex terrain are more complicated than those for the wind speed. Wind direction is found to depend primarily on the geostrophic wind direction (D_g), the aspect angle (A_{sp}) and the up-canyon direction, of which the values for the mountain and valley stations are listed in Table 5:

In view of the comparatively strong turbulence motion during the daytime, surface wind is more affected by the geostrophic wind than the topography of the terrain, whereas in the evening and early morning it is more affected by the terrain topography than the geostrophic wind. For this reason, statistical models for wind direction are classified according to time periods 3, 4, 5, 6 (daytime) and time periods 7, 8, 1, 2 (evening and early morning). The wind direction models for the mountain and valley stations are as follows:

Mountain Station SRII

If $D_g > A_{sp}$,

$$D = D_g - 0.32(D_g + \alpha_o - A_{sp}),$$

for periods 3, 4, 5, 6

$$D = D_{uc} - 0.50(D_g - A_{sp}),$$

for periods 7, 8, 1, 2

If $D_g \leq A_{sp}$,

$$D = D_g + 0.32(D_g + \alpha_o - A_{sp}),$$

for all periods

If $D_g \leq 180^\circ$,

$$D = D_g + \alpha_o$$

for all periods

Valley Station SRI5

If $D_g > A_{sp} - \alpha_o$,

$$D = D_g - 0.50(D_g + \alpha_o - A_{sp}),$$

for periods 3, 4, 5, 6

$$D = D_{uc}$$

for periods 7, 8, 1, 2

If $D_g \leq A_{sp} - \alpha_o$,

$$D = D_g + 0.50(D_g + \alpha_o - A_{sp}),$$

for periods 3, 4, 5, 6

$$D = D_{uc} - 0.80(D_g + \alpha_o - A_{sp}),$$

for periods 7, 8, 1, 2

If $D_g < 180^\circ - \alpha_o$,

$$D = D_{uc}$$

for all periods

Comparing the observed and calculated wind directions, we have found that the difference (Δ) between the calculated and observed wind direction having the following percentages for the mountain and valley stations:

Mountain Station SRI1

$\Delta < 20^\circ$	47.8%
$20^\circ \leq \Delta < 40^\circ$	14.0%
$40^\circ \leq \Delta < 60^\circ$	5.8%
$60^\circ \leq \Delta < 80^\circ$	3.0%
$\Delta < 80^\circ$	29.0%

Valley Station SRI5

$\Delta < 20^\circ$	52.2%
$20^\circ \leq \Delta < 40^\circ$	16.9%
$40^\circ \leq \Delta < 60^\circ$	6.6%
$60^\circ \leq \Delta < 80^\circ$	4.3%
$\Delta < 80^\circ$	20.0%

4. CONCLUSIONS

Statistical models for surface wind predictions at a mountain and a valley station near Anderson Creek, California, have been constructed. It is found that the surface wind speed depends primarily on the slope, valley and geostrophic winds. The correlations between the calculated and observed surface wind speeds is found high for all time periods of the day and night.

Because of the variability of wind direction, which is greatly affected by topography, geostrophic wind and turbulent motion, is generally larger than that of the surface wind speed, statistical models for wind direction are more complicated than those for the wind speed. It is found that wind direction depends primarily on the geostrophic wind direction, aspect angle of the topography, up-canyon direction and cross isobaric angle in the boundary layer.

5. ACKNOWLEDGMENT

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TABLE 1. Values of Fourier coefficients for the wind speeds.

	$\bar{V}_v$	A_1	B_1	A_2	B_2	A_3	B_3
Mountain Station	-0.51	-5.10	-1.60	1.80	0.90	0.57	0.31
Valley Station	-0.59	-2.10	-0.30	0.12	-0.04	0.25	0.23

TABLE 2. Values of coefficients a_1 , a_2 , a_3 , a_4 and a_5 for the mountain station at various time periods.

Time Period	1	2	3	4	5	6	7	8
a_1	115.71	159.32	33.23	91.98	7.06	-85.82	-114.96	-27.12
a_2	-50.71	-26.30	-46.69	-16.02	9.07	-177.17	-8.04	-10.49
a_3	11.44	38.26	133.39	68.90	1.34	-36.89	-48.89	-29.86
a_4	-0.02	-0.04	-0.22	-0.09	-0.01	0.04	0.08	0.04
a_5	-6.67	-19.74	-67.93	-35.29	-1.71	18.46	23.93	14.13

TABLE 3. Values of coefficients a_1 , a_2 , a_3 , a_4 and a_5 for the valley station at various time periods.

Time Period	1	2	3	4	5	6	7	8
a_1	-35.59	-7.57	73.34	-33.05	166.64	-53.78	3.26	-32.59
a_2	28.97	5.68	-19.88	16.46	-63.87	72.71	-7.59	19.98
a_3	-12.42	5.88	1.31	-11.22	5.03	-15.85	3.88	4.30
a_4	-0.03	-0.01	0.01	0.07	-0.04	0.07	-0.01	-0.01
a_5	6.78	-3.39	-0.18	6.30	-3.33	8.28	-2.26	-2.26

TABLE 4. Correlation coefficients between the observed and predicted wind speeds.

Time Period	1	2	3	4	5	6	7	8
Mountain	0.830	0.704	0.872	0.978	0.977	0.952	0.874	0.821
Valley	0.921	0.987	0.874	0.983	0.975	0.958	0.854	0.907

TABLE 5. Terrain constants for the mountain and valley stations.

	D_{uc} (upcanyon direction)	A_{sp} (aspect angle)	S (slope of terrain)
Mountain	45°	200°	4.5°
Valley	-15°	230°	5.4°

FIGURE 1. Topographic configuration near Anderson Creek, California.

MOUNTAIN

FIGURE 2. Observed and predicted wind speeds at the mountain station.

VALLEY

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FIGURE 3. Observed and predicted wind speeds at the valley station.

A NONSTATIONARY NOCTURNAL DRAINAGE FLOW MODEL

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ABSTRACT

The evolution and structure of the steady state of an idealized nocturnal drainage flow over a large uniformly-sloping surface are studied using a nonstationary model with a height-dependent eddy diffusivity profile and a specified surface cooling rate. The predicted mean velocity and temperature profiles are compared with Prandtl's stationary analytical solutions based on the assumption of a constant eddy diffusivity in the drainage layer. The effects of important physical parameters, such as the slope angle, surface cooling, atmospheric stability, and surface roughness, on the steady drainage flow are investigated.

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1. INTRODUCTION

The nocturnal drainage flow considered in this paper is a gravitationally induced local wind on a slope caused by the radiative cooling of the surface. This well-known phenomenon, usually referred to as the katabatic or downslope wind, may also occur in the daytime over ice or snow covered slopes. The local winds have a perceptible effect on climate, crops, and human welfare, especially if a source of pollution is involved. Hence, an understanding of the physical processes and structure of these flows is important.

Several analytical and numerical model studies of the drainage flow have been reported in the literature. Summaries of this work and related references can be found in Defant (1951), Thyer (1966), Gutman (1972), and Manins and Sawford (1979). In a pioneering theory of the steady slope wind over a large homogeneous sloping plane, Prandtl (1942) derived simple analytical expressions for the mean wind speed and potential temperature profiles as functions of the distance normal to the slope, surface temperature deviation from a reference state value, and a scale height of the flow. He assumed the flow to be stationary, the eddy diffusivities of momentum and heat (K_m and K_h) to be constant and equal, and the slope angle to be small. Defant (1949) extended Prandtl's theory to include the oscillatory nature of the slope wind and large slope angles and, using $K_h/K_m = 1.4$, obtained good agreement between the predicted and observed wind profiles on a steep slope in the Innsbruck range. Lettau (1966) also found a satisfactory fit between Prandtl's theory and the observed wind and temperature distributions in a katabatic flow on the south polar plateau.

Nevertheless, the height dependence of the eddy diffusivities was not taken into account in Prandtl's theory, and quantitative comparison then reduced to a determination of a constant K coefficient for an arbitrarily estimated value of the surface temperature deviation. This paper describes a nonstationary nocturnal drainage flow model using a variable eddy diffusivity profile based on considerations of the flow structure, surface cooling rate, ambient atmospheric conditions, and the surface characteristics. Numerical results on the evolution and structure of the steady drainage flow are presented. The effects of important physical parameters, such as the slope angle, surface cooling, atmospheric stability, and surface roughness, on the drainage flow are investigated.

2. THE MODEL

We consider the evolution of an idealized drainage flow over a large homogenous plane surface, sloping uniformly at an angle β to the horizontal (x, y) plane. With the origin O located somewhere on the surface, we adopt a slope-aligned rectangular coordinate system (s, y, n) such that the s axis is oriented along the fall-line vector, y axis in the lateral direction, and n axis normal to the surface (Fig. 1). Because of the homogeneity of surface roughness and temperature, and the flow symmetry in this large plane problem, all components of motion are assumed to depend only on the normal height above the surface (n) and time (t). The Coriolis effects are negligible except for very small slope angles. Then the governing equations for the katabatic flow in the absence of an external wind can be written (Rao and Snodgrass, 1979) as

$$\partial U / \partial t = -(g/\theta_0) \theta' \sin \beta + \partial (K_m \partial U / \partial n) / \partial n \quad (1)$$

$$\partial \theta / \partial t = U \gamma \sin \beta + \partial (K_h \partial \theta / \partial n) / \partial n$$

Here, U is the mean flow velocity parallel to the s axis, $\theta'(n) = \theta - \theta_0$ is the deviation of the mean potential temperature θ from an undisturbed atmospheric reference state distribution $\theta_0(z)$, defined such that $\partial \theta_0 / \partial t = 0$ and $\partial \theta_0 / \partial z = \gamma = \text{constant} > 0$, and g is the acceleration due to gravity. The first equation states that the downslope acceleration is proportional to the imbalance between the drainage and turbulent friction forces. The second equation indicates that the cooling rate of the drainage layer is equal to the imbalance between the turbulent heat flux divergence and rate of downslope-warming due to the stable stratification of the ambient atmosphere.

For the flow over a sloping plane with β not less than several degrees and a characteristic value of the surface temperature deviation of at least a few degrees in magnitude, the gradients of mean temperature deviations normal to the slope are typically an order of magnitude larger than the temperature gradients in the undisturbed atmosphere and, hence, the latter can be neglected (Gutman, 1972). This simplifies the mean temperature equation as

$$\partial \theta' / \partial t = U \gamma \sin \beta + \partial (K_h \partial \theta' / \partial n) / \partial n \quad (2)$$

The eddy diffusivities of momentum and heat are calculated as follows:

$$K_{m,h} = \ell_{m,h} (aE)^{\frac{1}{2}}, \quad \ell_{m,h} = kn/\phi_{m,h} \quad (3a)$$

where $\ell_{m,h}$ are the turbulent mixing lengths, E is the turbulent kinetic energy (TKE), $k = 0.35$ is von Kármán's constant, and $a=0.2$ is a proportionality constant (Delage, 1974; Nieuwstadt and Driedonks, 1979). The nondimensional wind shear (ϕ_m) and potential temperature gradient (ϕ_h) are specified by the expressions due to Businger et al. (1971) for the stable case; E is calculated from the TKE equation,

$$\begin{aligned} \partial E / \partial t = & \partial (K_m \partial E / \partial n) / \partial n + K_m (\partial U / \partial n)^2 - (g / \theta_o) K_h (\partial \theta' / \partial n + \gamma \cos \beta) \\ & - b(aE)^2 / K_m, \end{aligned} \quad (3b)$$

where $b=1-R_f$, and R_f is the local flux Richardson number computed directly from the model such that $0 \leq R_f \leq R_{fcr} = 0.21$. Numerical solutions of the closed equation set (1) to (3), subject to the boundary and initial conditions given below, were obtained using a Dufort-Frankel explicit finite difference scheme. A logarithmic transformation of the vertical coordinate was used to obtain a fine computational mesh near the ground.

At $n=z_o$, the aerodynamic surface roughness, the boundary conditions are

$$U = 0, \quad \theta' = -\Delta_o, \quad \partial E / \partial n = 0, \quad (4a)$$

where $\Delta_o(t)$ is the magnitude of the surface temperature deviation, defined such that $\Delta_o(0) = 0$. The flux-profile relations of Businger et al. (1971) for the stable case are used to calculate the turbulent

fluxes in terms of mean field variables at the lower boundary.

At $n = H > h$, where $h(t)$ is the thickness of the drainage layer, not known a priori but determined from the solution of the problem, we specify

$$U = \theta' = E = 0 \quad (4b)$$

For a pure katabatic flow in the absence of an external wind, Eq. (4b) also describes the conditions at $t = 0$ since the air is initially at rest. For $t > 0$, the air near the surface cools due to the rapid radiative heat loss from the surface on a clear night. A flow develops when the cooled air layer is heavy enough to overcome friction and slide down the slope. Presumably this surface drainage flow will be laminar in the early stage of its development, but a transition to turbulence occurs within a short time depending on the flow Reynolds number. As the surface continues to cool, the wind speed and the turbulent mixing increase, and the mean quantities of the atmospheric drainage flow are determined from Eq. (1) in which the molecular contributions to the momentum and heat fluxes are neglected. Even if the molecular terms were included in the formulations, the model integration cannot be started from rest due to numerical instabilities encountered in the transition from the laminar to the turbulent regime.

In order to obtain a numerical solution of the turbulent drainage flow described by Eqs. (1) to (4), therefore, the initial conditions on U , θ' and E must be given at a time $t = t_0 > 0$. These initial distributions may be approximated, in principle, by any compatible set of profiles that satisfy the boundary conditions (4a) and (4b), as well as the flux-profile relations at the lower boundary. Due to their simplicity and

convenience, Prandtl's analytical solutions (see Section 3.2) were used to specify the approximate initial profiles following a procedure discussed by Rao and Snodgrass (1979). The governing equations were then numerically integrated in time, holding the boundary conditions constant, until the equilibrium distributions consistent with our height-dependent eddy diffusivity formulations evolved; these distributions were then used as the correct initial conditions at $t=t_0$. This procedure ensures that the model results, presented below, are independent of the approximate initial conditions specified to start the numerical integration.

3. RESULTS AND DISCUSSION

3.1 Evolution and Structure of Steady State Flow

The evolution of the mean velocity and potential temperature profiles, and the corresponding eddy diffusivity profiles, in a nocturnal drainage flow are shown in Figs. 2(a), (b), and (c), respectively. The flow parameters used in the computations are $\beta = 10^\circ$, $\gamma = 0.006 \text{ C m}^{-1}$, $z_0 = 0.1\text{m}$, and a surface cooling rate of 2 C hr^{-1} for $0 < t < 1 \text{ hr}$. For $t \geq 1 \text{ hr}$, the surface temperature was kept constant such that $\Delta_0 = 2^\circ\text{C}$. Since the surface temperature variation specified here was rather arbitrary, the time t shown in the figures should be interpreted only as an indicator of the corresponding surface temperature-deficit (Δ_0), and not the physical time per se. A striking feature of the slope wind profile is the occurrence of a prominent velocity maximum ($U_{\max}$) near the surface. The magnitude and location of this low-level drainage jet vary as a function of $\Delta_0(t)$, among other factors, as shown in Fig. 2(a). The

small positive deviations at the top of the temperature profiles (Fig. 2b) can be explained as follows. Due to turbulent friction, the cooled sinking air parcels also drag down the adjacent non-cooled air layers. During their descent, the latter become warmer than the neighboring layers which are not set into motion. These layers, upon being warmed by the turbulent transfer of heat, start to ascend and form a weak counterflow at the top of the drainage layer (Fig. 2a).

A stationary solution of the drainage flow described by Eqs. (1) to (4) can be obtained when Δ_0 approaches a constant value. In this example, with Δ_0 held constant at 2°C for $t \geq 1$ hr, a steady state, with a peak velocity of 1.5 m s^{-1} , was attained in a few hours as indicated by the invariancy of profiles at $t = 6$ and 11 hrs which coincide exactly in Fig. 2.

The steady state profiles of the turbulent kinetic energy E , shear stress $uw = -K_m \partial U / \partial n$, and heat flux $w\theta = -K_h \partial \theta' / \partial n$ are shown in Figs. 3(a), (b), and (c), respectively. The kink in the TKE profile corresponds to the maximum in the velocity profile, where the local shear production is zero. At this level, the non-zero energy is only due to the turbulent transport from the surface. The shear stress (Fig. 3b) is negative near the ground, goes through zero at the U_{max} - level, and remains positive in the outer layer. A positive maximum in uw occurs at a level corresponding to the inflection point in the velocity profile where $\partial^2 U / \partial n^2 = 0$. The heat flux (Fig. 3c) is negative in the drainage flow with a maximum near the surface. The slight positive heat flux at the top is due to the positive temperature deviations discussed above. Based on the turbulence profiles, the

thickness h of the stationary drainage layer in this problem is 32 m.

3.2 Comparison with Prandtl's Theory

Prandtl's analytical expressions (see Defant, 1951) for the mean wind and temperature deviation profiles in the stationary slope wind can be written as

$$U(\xi) = U_c e^{\xi} \sin \xi, \quad \theta'(\xi) = -\Delta_0 e^{\xi} \cos \xi, \quad (5a)$$

where

$$\xi = (n - z_0)/\ell, \\ \ell = (2K_m/N \sin \beta)^{1/2} \alpha_k^{1/2}, \quad U_c = (\Delta_0 N/\gamma) \alpha_k^{1/2}. \quad (5b)$$

In the above, K_m is a constant eddy diffusivity, $\alpha_k = K_h/K_m$ is a constant of order unity, $N = (g\gamma/\theta_0)^{1/2}$ is the Brunt-Väisälä frequency in the stable ambient atmosphere, and ℓ and U_c , respectively, are the characteristic height and velocity of the drainage flow. The original Prandtl solutions, which do not explicitly depend on the surface roughness parameter, are slightly modified here, by including z_0 in the definition of ξ in Eq. (5b), to be consistent with our formulations. Equations (5) indicate that the slope wind is effectively confined to the layer $z_0 \leq n \leq \pi\ell$, and it reaches its first and absolute maximum, $U_{\max} = 0.3224U_c$, at a normal height $n_{\max} = \pi\ell/4 + z_0$ where $\partial U/\partial n = 0$. Therefore, U_c and ℓ , and hence the mean wind and temperature distributions from Eqs. (5), can be determined if $U_{\max}$ and $n_{\max}$ are known from a field or a numerical experiment.

Figures 4(a), (b), and (c) show the comparison of U , θ' , and K_m profiles, respectively, between the present model calculations

and Prandtl's theory. The latter, using $\alpha_k = 1.35$, underpredicts the drainage layer thickness and yields smaller velocities at all levels, while slightly overpredicting θ' in the surface layer. However, the general profile shape and behavior are similar in both models. This suggests that Eqs. (5) can be used to provide a fair approximation to the velocity and temperature profiles in a katabatic flow if the peak jet velocity and its height are independently determined from field observations. Earlier work by Defant (1949) and Lettau (1966) support this conclusion.

3.3 Sensitivity Analyses

In order to study the relative effects of important physical parameters on the steady nocturnal drainage flow, each parameter was varied while keeping the others constant in the computations. The parameters of the control run are $\beta = 10^\circ$, $\Delta_0 = 4^\circ\text{C}$, $\gamma = 0.006\text{C m}^{-1}$, and $z_0 = 0.1\text{ m}$. Figures 5(a) and (b), respectively, show the steady state profiles of U and θ' for three different slope angles. As β increases, the drainage velocity at a given height increases near the surface and decreases in the outer layer. However, the maximum velocity ($U_{\max}$) is not very sensitive to the slope angle; the physical explanation is that while it is more difficult for the air to sink along a steeper slope, the drainage force, which is proportional to $\sin \beta$, is also larger. As the slope increases, comparatively less cooling (Fig. 5b) of the air layer is required to produce the same drainage force. It can be seen that the drainage layer becomes shallower and $n_{\max}$ decreases as the slope increases.

Figures 6(a) and (b) show the steady state U and θ' profiles for three different values of the surface temperature-deficit (Δ_0). As the latter increases, U and θ' increase at all levels; $U_{\max}$, $n_{\max}$, and h also increase. The variation of the maximum drainage velocity is directly proportional to the surface cooling. The effects of atmospheric stability on U and θ' profiles in the steady drainage flow were also investigated (not shown) using three different values of γ between adiabatic and isothermal conditions. The results indicate that the more stable the ambient atmosphere, the less are the velocity and cooling in the drainage flow; the boundary layer thickness also is smaller.

The effects of the surface roughness on the equilibrium profiles of U , θ' , and K_m in the drainage flow are shown in Figs. 7(a), (b), and (c). As z_0 decreases, the turbulent friction force decreases; however, the cooling in the layer and the drainage force also decrease since all other flow parameters including the surface cooling (Δ_0) were kept constant in this study. Thus, as z_0 decreases, $U_{\max}$, $n_{\max}$, and h decrease as shown.

4. CONCLUSIONS

The evolution of an idealized nocturnal drainage flow over a large homogeneous uniformly-sloping surface has been investigated by a nonstationary model using a height-dependent eddy diffusivity profile and a specified surface cooling rate. For a constant surface temperature-deficit, the flow approaches stationarity. Though it is unlikely that long periods of this steady state occur in reality when the ground temperature is determined by the surface energy budget (Brost and Wyngaard, 1978), the model is useful in understanding

the equilibrium structure of the katabatic flow. The evolution of the drainage flow can be predicted if the surface temperature or heat flux variation, among other parameters, is determined from field experiments. Due to lack of suitable data, this numerical model has not yet been tested directly against observations. However, the satisfactory fit of Prandtl's theory to the field data reported by Defant (1949) and Lettau (1966), and the mean velocity and density profiles obtained in the laboratory simulations of Ellison and Turner (1959) indirectly support the model predictions.

The simplifying assumptions that led to this one-dimensional model do not allow for the explicit accounting of the effects of the interfacial entrainment of ambient air into the drainage flow that are known to be important (Ellison and Turner, 1959; Manins and Sawford, 1979) for slopes of limited extent. Nevertheless, this model should yield satisfactory results for large sections of mountain slopes (Gutman, 1972), away from the ridges and valleys.

This drainage flow model is applicable only for relatively calm or light synoptic wind conditions. The extension of the model to account for the ambient winds and Coriolis effects, the conditions for the existence of a steady state, and the stationary analytical solutions are discussed by Rao and Snodgrass (1979).

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FIGURE 1. Schematic diagram of the slope-aligned coordinate system.

FIGURE 2a. Predicted evolution of mean velocity. The profiles for $t=6$ and 11 hrs., which coincide exactly, indicate the establishment of a steady state. See text for details and flow parameters.

FIGURE 2b. Predicted evolution of mean potential temperature. The profiles for $t=6$ and 11 hrs., which coincide exactly, indicate the establishment of a steady state. See text for details and flow parameters.

FIGURE 2c. Predicted evolution of eddy diffusivity of momentum profiles in the drainage flow. The profiles for $t=6$ and 11 hrs., which coincide exactly, indicate the establishment of a steady state. See text for details and flow parameters.

FIGURE 3a. Steady state profiles of the turbulent kinetic energy.

FIGURE 3b. Steady state profiles of the turbulent shear stress.

FIGURE 3c. Steady state profiles of the turbulent heat flux in the drainage flow ($\beta = 10^\circ$, $\Delta_0 = 2^\circ\text{C}$, $\gamma = 0.006 \text{ C m}^{-1}$, $z_0 = 0.1 \text{ m}$).

FIGURE 4a. Comparison of the steady state mean velocity.

FIGURE 4b. Comparison of the steady state mean potential temperature deviation.

FIGURE 4c. Comparison of the steady state eddy diffusivity of momentum profiles predicted by the present model (—) and Prandtl's theory (-----).

FIGURE 5a. Effects of the variation of the slope angle β on the steady state U .

FIGURE 5b. Effects of the variation of the slope angle β on the steady state θ' profiles in the drainage flow ($\Delta_0 = 4^\circ\text{C}$, $\gamma = 0.006 \text{ C m}^{-1}$, $z_0 = 0.1 \text{ m}$).

FIGURE 6a. Effects of the variation of the surface temperature-deficit Δ_0 on the steady state U .

FIGURE 6b. Effects of the variation of the surface temperature-deficit Δ_0 on the steady state Θ' profiles in the drainage flow ($\beta = 10^\circ$, $\gamma = 0.006 \text{ C m}^{-1}$, $z_0 = 0.1 \text{ m}$).

FIGURE 7a. Effects of the variation of the surface roughness parameter z_0 on the steady state U .

FIGURE 7b. Effects of the variation of the surface roughness parameter z_0 on the steady state θ' .

FIGURE 7c. Effects of the variation of the surface roughness parameter z_0 on the steady state K_m profiles in the drainage flow ($\beta = 10^\circ$, $\Delta_0 = 4^\circ\text{C}$, $\gamma = 0.006 \text{ C m}^{-1}$).

A NUMERICAL SIMULATION OF NOCTURNAL DRAINAGE FLOW

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Abstract

A three-dimensional mesoscale atmospheric model, previously used to simulate airflow over cooling ponds and hypothetical, simplified mountains, has been modified to simulate development of nocturnal drainage flow in realistic terrain, part of the California Geysers area. Turbulent fluxes appearing in the governing equations for mean variables are computed from simplified second-moment turbulence-closure equations where only a turbulence energy equation and a master-length scale equation are solved prognostically. Qualitative simulation of drainage flow, horizontal convergence in the valley, and resulting vertical motions are satisfactory. Simulated wind speeds near the surface are greater than observed values, possibly due to inadequate resolution of observations or to a lack of consideration of the drag associated with canopy flow.

1. Introduction

One of the objectives of the Atmospheric Study in Complex Terrain (ASCOT) program of the U. S. Department of Energy, is to improve modeling of transport and diffusion of energy-related pollutants over terrain typical of areas of energy development. An initial ASCOT field experiment to investigate nocturnal drainage flow, conducted during summer 1979 over an area designated as the California Known Geothermal Resources Area (KGRA), provides data which can be compared with model simulations of wind flows.

Analytical solutions for density driven slope flow have been obtained under special conditions in which various simplifications can be assumed (e.g., see Sutton, 1953). The most critical assumption in the theory is one-dimensional flow, i.e., the flow is assumed to be uniform except in the direction perpendicular to the surface. Eddy viscosity and eddy diffusivity are also assumed to be constant. Despite these simplifications, wind profiles predicted by the theory agree qualitatively with field data such as that obtained at Innsbruck, Austria (Sutton, 1953). Analytical solutions for a two-dimensional case in which quantities vary tangentially as well as normally to the slope have been obtained for mountain lee-wave flow. However, velocities in layers close to the surface, such as in the case of nocturnal drainage flow, are difficult to obtain, since the boundary conditions at the surface are quite

complex. Thus, analytical solutions are often replaced by numerical solutions obtained with finite-difference or finite-element methods.

The complex topography in the Geysers geothermal area results in highly variable wind and turbulence fields. Vertical motion, shear-generated turbulence and meandering of horizontal winds in the Geysers area are all important for realistic simulations of dispersion of hydrogen sulfide released into the atmosphere during conversion of geothermal energy into electric power. Three-dimensional mesoscale models that consider the effects of topography have been proposed (e.g., Anthes and Warner, 1974; Mahrer and Pielke, 1975; Nickerson, 1979). However, none of the models have been applied to an area whose horizontal scale is as small as that studied here, and most existing models use a simple expression for eddy viscosity (e.g. O'Brian, 1970) for parameterization of turbulence. O'Brian's formula requires specification of the boundary layer height. Estimation of boundary layer heights over rugged terrain is difficult, particularly when the atmosphere is stably stratified.

In this study, a three-dimensional, simplified second-moment turbulence-closure model, previously used to simulate airflow over Gaussian-shaped mountains (Yamada, 1978b) and a large cooling pond (Yamada, 1979), has been modified to simulate nocturnal drainage flow over the Geysers area. Results are compared with the wind

data collected during the 1979 field experiment, briefly reviewed in the following section.

2. Initial ASCOT field experiment

The first preliminary field experiment under ASCOT was conducted in July 1979 over an area approximately 10 km x 7 km (Fig. 1) in the California KGRA, approximately 130 km north of San Francisco. The area is surrounded by three mountains, Boggs Mountain (elevation ~ 1000 m) to the north, Cobb Mountain (~ 1300 m) to the northeast and Pine Mountain (~ 1000 m) to the southwest, which form a valley opening toward the east-southeast. Surface vegetation is heterogeneous, varying from dense forest to almost bare soil. However, at present, no quantitative information on distribution of surface vegetation is available. Five National Laboratories and other institutions participated in the experiment, which lasted approximately two weeks. Vertical profiles of mean wind, mean temperature, and mean water vapor up to 525 m above the ground were measured hourly at two sites by tethered sondes from sunset to sunrise. Additional wind profiles were obtained with a doppler acoustic sounder and pibals. Winds averaged over a 600 m path were recorded by a laser anemometer. Winds at 10 m above the ground were measured hourly at seven towers deployed in the area. These data are used in Section 7 to evaluate model results.

3. Model equations

For simulations of flows over nonhomogeneous surface boundaries, a terrain following vertical coordinate system is often used in order to simplify the treatment of boundary conditions.

Here the transformation used by Mahrer and Pielke (1975) is adopted:

$$z^* = \bar{H} (z - z_g(x,y)) / (H(x,y,t) - z_g(x,y)), \quad (1)$$

where z^* and z are the transformed and Cartesian vertical coordinates, respectively; $z_g(x,y)$ is ground elevation; $H(x,y,t)$ is the material surface top of the model and $\bar{H}$ is the initial value of H . It is noted that z^* is directed vertically and may not be perpendicular to the ground.

The governing equations, following the coordinate transformation, are*

$$\begin{aligned} DU/Dt = & f(V - V_g) + g(\bar{H} - z^*)(\bar{H})^{-1} [1 - 2 \langle \Theta_v \rangle \Theta_v^{-1}] \partial z_g / \partial x \\ & + g[2 \langle \Theta_v \rangle \langle \Theta_v(\bar{H}) \rangle^{-1} - (2 \langle \Theta_v \rangle \Theta_v^{-1} - 1) z^*(\bar{H})^{-1} - 1] \cdot \partial H / \partial x \\ & + \partial(K_x \partial U / \partial x) / \partial x + \partial(K_{xy} \partial U / \partial y) / \partial y + \bar{H}(H - z_g)^{-1} \partial(-\overline{uw}) / \partial z^*, \end{aligned} \quad (2)$$

$$\begin{aligned} DV/Dt = & -f(U - U_g) \\ & + g(\bar{H} - z^*)(\bar{H})^{-1} [1 - 2 \langle \Theta_v \rangle \Theta_v^{-1}] \partial z_g / \partial y \\ & + g[2 \langle \Theta_v \rangle \langle \Theta_v(\bar{H}) \rangle^{-1} - (2 \langle \Theta_v \rangle \Theta_v^{-1} - 1) z^*(\bar{H})^{-1} - 1] \partial H / \partial y \\ & + \partial(K_{xy} \partial V / \partial x) / \partial x + \partial(K_y \partial V / \partial y) / \partial y + \bar{H}(H - z_g)^{-1} \partial(-\overline{vw}) / \partial z^*, \end{aligned} \quad (3)$$

$$\begin{aligned} \partial U / \partial x + \partial V / \partial y + \partial W^* / \partial z^* - (H - z_g)^{-1} (U \partial z_g / \partial x \\ + V \partial z_g / \partial y) + (H - z_g)^{-1} (\partial H / \partial t + U \partial H / \partial x + V \partial H / \partial y) = 0 \end{aligned} \quad (4)$$

* Derivations of (2), (3), (8), and (9), and the assumptions used are found in the Appendix. A list of symbols is also provided in the Appendix.

where

$$W^* \equiv \bar{H}(H - z_g)^{-1} W - z^*(H - z_g)^{-1} (\partial H / \partial t + U \partial H / \partial x + V \partial H / \partial y) + (z^* - \bar{H})(H - z_g)^{-1} (U \partial z_g / \partial x + V \partial z_g / \partial y), \quad (5)$$

and

$$D(\) / Dt \equiv \partial(\) / \partial t + U \partial(\) / \partial x + V \partial(\) / \partial y + W \partial(\) / \partial z = \partial(\) / \partial t + U \partial(\) / \partial x + V \partial(\) / \partial y + W^* \partial(\) / \partial z^*. \quad (6)$$

In the above expressions $\langle \rangle$ indicates an average over a horizontal surface in the terrain-following coordinate system.

A prognostic equation for H may be obtained by integrating the equation of continuity (4) from the surface to $\bar{H}$ (Mahrer and Pielke, 1975). However, in this study, a simple relationship

$$H = \bar{H} + z_g(x, y), \quad (7)$$

is assumed.

The geostrophic wind components in (2) and (3) are given as

$$\begin{aligned} fU_g &= fU_g(\bar{H}) \cdot \langle \Theta_v \rangle \cdot \langle \Theta_v(\bar{H}) \rangle^{-1} \\ &+ g(H - z_g)(\bar{H})^{-1} \int_{z_*}^{\bar{H}} \langle \Theta_v \rangle^{-1} \cdot \partial \Delta \Theta_v / \partial y \cdot dz' \\ &+ g(\bar{H})^{-1} (\partial H / \partial y - \partial z_g / \partial y) \int_{z_*}^{\bar{H}} \Delta \Theta_v dz' \end{aligned} \quad (8)$$

and

$$\begin{aligned} fV_g &\equiv fV_g(\bar{H}) \cdot \langle \Theta_v \rangle \cdot \langle \Theta_v(\bar{H}) \rangle^{-1} \\ &- g(H - z_g)(\bar{H})^{-1} \int_{z_*}^{\bar{H}} \langle \Theta_v \rangle^{-1} \partial \Delta \Theta_v / \partial x \cdot dz' \\ &- g(\bar{H})^{-1} (\partial H / \partial x - \partial z_g / \partial x) \int_{z_*}^{\bar{H}} \Delta \Theta_v dz', \end{aligned} \quad (9)$$

where $\Delta \Theta_v = \Theta_v - \langle \Theta_v \rangle$, and abbreviated symbols $U_g(\bar{H}) = U_g(x, y, \bar{H}, t)$, $V_g(\bar{H}) = V_h(x, y, \bar{H}, t)$ etc. are used. For deriving (8) and (9), the hydrostatic equation, the Boussinesq approximation, and boundary conditions at $z^* = \bar{H}$; $DU/Dt = DV/Dt = 0$, $U = U_g(\bar{H})$, and $V = V_g(\bar{H})$ are used (see the Appendix for detailed discussions).

Although simulation of clouds is not yet a major interest in this study, special attention should be given to the somewhat unconventional approach employed here for treatment of condensation and evaporation processes. The conservative variables, appropriate to the cases where phase changes occur but do not produce rain, are the liquid-water potential temperature (Betts, 1973)

$$\Theta_\ell \equiv \Theta - (\Theta/T)(L/c_p)Q_\ell, \quad (10)$$

and the mixing ratio of the total water content

$$Q_w = Q_v + Q_\ell \quad (11)$$

where Θ is the potential temperature, and Q_v and Q_ℓ are the mixing ratios of water vapor and liquid water content, respectively. Detailed discussions, including derivations of the conservation equations of Θ_ℓ and Q_w , are given in Yamada (1978a). The resulting equations, following the coordinate transformation, are

$$\begin{aligned} D\Theta_\ell/Dt = & \partial(K_x \partial \Theta_\ell / \partial x) / \partial x + \partial(K_y \partial \Theta_\ell / \partial y) / \partial y \\ & + \bar{H}(H - z_g)^{-1} \partial(-\overline{w\theta_\ell}) / \partial z^*, \end{aligned} \quad (12)$$

and

$$\begin{aligned} DQ_w/Dt = & \partial(K_x \partial Q_w / \partial x) / \partial x + \partial(K_y \partial Q_w / \partial y) / \partial y \\ & + \bar{H}(H - z_g)^{-1} \partial(-\overline{wq_w}) / \partial z^*. \end{aligned} \quad (13)$$

The vertical turbulent fluxes $\overline{uw}$, $\overline{vw}$, $\overline{w\theta}_\ell$ and $\overline{wq}_w$ in (2), (3), (12) and (13) are computed from a set of simplified second-moment turbulence closure equations in which only a turbulence energy equation and a master length-scale equation (Mellor and Yamada, 1977) are solved prognostically. The turbulence energy equation is given by

$$\begin{aligned} D(q^2/2)/Dt = & \partial\{K_x \partial(q^2/2)/\partial x\}/\partial x + \partial\{K_y \partial(q^2/2)/\partial y\}/\partial y \\ & + \{\bar{H}/(H - z_g)\}^2 \partial\{q\ell\tilde{S}_q \partial(q^2/2)/\partial z^*\}/\partial z^* \\ & - \bar{H}(H - z_g)^{-1} (\overline{uw} \partial U/\partial z^* + \overline{vw} \partial V/\partial z^*) \\ & + \beta g \overline{w\theta}_v - q^3 (B_1 \ell)^{-1}, \end{aligned} \quad (14)$$

and the master length-scale equation is given by

$$\begin{aligned} D(q^2 \ell)/Dt = & \partial\{K_x \partial(q^2 \ell)/\partial x\}/\partial x + \partial\{K_y \partial(q^2 \ell)/\partial y\}/\partial y \\ & + \{\bar{H}/(H - z_g)\}^2 \partial\{q\ell\tilde{S}_\ell \partial(q^2 \ell)/\partial z^*\}/\partial z^* \\ & + \ell E_1 [\bar{H}(H - z_g)^{-1} (-\overline{uw} \partial U/\partial z^* - \overline{vw} \partial V/\partial z^*) + \\ & + \beta g \overline{w\theta}_v] - q^3 B_1^{-1} [1 + E_2 (\ell/kz)^2]. \end{aligned} \quad (15)$$

The remaining turbulent second moments are obtained from a set of diagnostic equations; the final expressions for turbulent fluxes $\overline{uw}$, $\overline{vw}$, $\overline{w\theta}_\ell$, $\overline{wq}_w$ and $\overline{w\theta}_v$ are given elsewhere (Yamada, 1978a, b, and 1979). Thus, they are not repeated here.

In the previous study (Yamada, 1978b), the temperature at screen height was specified and solutions in approximately steady state were obtained. However, in this study, the surface temperature

is determined by solving a heat conduction equation for soil layers. Therefore, the boundary conditions are discussed in detail in the following section.

4. Boundary conditions

Surface boundary conditions for (2), (3), and (12) to (15) are constructed from the empirical formulas by Dyer and Hicks (1970) for the nondimensional wind and temperature profiles (see the Appendix). Strictly speaking, the formulas are valid only for horizontally homogeneous surfaces. It is assumed, however, that the same relations are fair approximations over nonhomogeneous terrain, provided that the formulas are applied sufficiently close to the surface. It should be noted that vegetation plays an active part in the apportionment of available heat energy between convective (sensible and latent) and conductive (into the soil) components. The technique discussed here is intended to address only the case of bare soil where the surface is conventionally characterized by roughness lengths. The complexity introduced by biological factors and drag forces due to tall trees (canopy flow) are beyond the scope of the present study. Use of the similarity formulas requires knowledge of the surface temperatures; a method to obtain the surface temperature is discussed below.

The temperature T_s in the soil layer is obtained by solving the heat conduction equation

$$\partial T_S / \partial t = \partial (k_S \partial T_S / \partial z_S) / \partial z_S, \quad (16)$$

where z_S is positive downward, and soil diffusivity k_S can be a function of soil moisture content. Appropriate boundary conditions for solution of (16) are the heat energy balance at the soil surface and specification of the soil temperature or soil heat flux at a depth dependent on the duration of the integration. The heat energy balance at the surface is given by

$$R_S + R_{L\downarrow} - R_{L\uparrow} = H_S + LE + G_S, \quad (17)$$

where R_S is the incoming solar radiation absorbed by the ground surface (R_S is zero in this study since only nocturnal cases are considered). $R_{L\downarrow}$ is the incoming longwave radiation, and $R_{L\uparrow}$ is the outgoing longwave radiation.

The surface heat flux H_S , latent heat flux LE and ground heat flux G_S are given by

$$H_S = - \rho_a c_p u_* T_* \quad (18)$$

$$LE = - \rho_a L u_* Q_* \quad (19)$$

and

$$G_S = - k_S \partial T_S / \partial z_S |_G \quad (20)$$

where ρ_a is the air density, u_* is the friction velocity, T_* is the temperature scale, Q_* is the water vapor scale, and the subscript G denotes the value at the ground surface. Substituting (18) to (20) into (17) we obtain

$$R_S + \epsilon R_{L\downarrow} - \epsilon \sigma T_G^4 = - \rho_a c_p u_* T_* (1 + B^{-1}) - k_S \partial T_S / \partial z_S |_G, \quad (21)$$

where the relation

$$R_L^\uparrow = \epsilon \sigma T_G^4 + (1 - \epsilon) R_L^\downarrow, \quad (22)$$

and Bowen ratio

$$B = H_S / LE \quad (23)$$

are used; ϵ is the emissivity of the surface and σ is the Stefan-Boltzman constant. Garratt and Hicks (1973) obtained a relationship between the surface temperature and air temperature at z_1 (in the surface layer):

$$(\overline{H}(z_1) - \overline{H}_G) T_*^{-1} = (P_r/k) [\ln\{(z_1+z_{ot})/z_o\} + \ln(z_o/z_{ot}) - \psi], \quad (24)$$

where P_r is the turbulence Prandtl number at neutral stability, k is the von Karman constant, z_o and z_{ot} are the roughness lengths for momentum and temperature, and ψ is the stability correction term of Panofsky (1963). A constant value of 0.1 m is assumed for z_o , and z_{ot} is obtained from a relationship $\ln(z_o/z_{ot}) = 2$ (Garratt and Hicks, 1973). Using (24) we can eliminate T_* from (21) to obtain

$$R_S + \epsilon R_L^\downarrow - \epsilon \sigma T_G^4 + m [H(z_1) - T_G (P_o/P_G)^{R/c_p}] + k_s \partial T_S / \partial z_s|_G = 0, \quad (25)$$

where

$$m = k \rho_a c_p u_* (1+B^{-1}) P_r^{-1} [\ln\{(z_1+z_{ot})/z_o\} + 2 - \psi]^{-1}. \quad (26)$$

P_o is a reference pressure (1000 mb) and P_G is the pressure at the surface. Equation (25) may be linearized by noting that $(T_G^{n+1} - T_G^n)/T_G^n \ll 1$, where the superscripts n and $n+1$ denote the

n and $(n + 1)$ th time steps of integration (a typical time increment used in integration is 1 min). After substitution into (25) of the approximation

$$(T_G^{n+1})^4 \approx 4(T_G^n)^3 T_G^{n+1} - 3(T_G^n)^4, \quad (27)$$

we obtain

$$\begin{aligned} [4\epsilon\sigma(T_G^n)^3 + m(P_o/P_G)^{R/c} p + k_s/\Delta z_s] T_G^{n+1} \\ = (k_s/\Delta z_s) T_s^{n+1}(1) + R_s + \epsilon R_L + 3\epsilon\sigma(T_G^n)^4 + m\theta^n(z_1), \end{aligned} \quad (28)$$

where the derivative $\partial T_s / \partial z_s|_G$ is replaced by a forward finite-difference approximation $(T_s^{n+1}(1) - T_G^{n+1})/\Delta z_s$, $T_s(1)$ is the soil temperature at the first grid level from the surface, and Δz_s is the distance between the surface and the first grid level in the soil layer.

Equation (16) is solved numerically in finite-difference form by Laasonen (Richtmyer and Morton, 1967, p. 189). By this method (16) reduces to $\underline{A} \underline{T}_s = \underline{B}$ where $\underline{A}$ is a tridiagonal matrix and $\underline{B}$ is a column vector. The solution is conveniently obtained by using the relation (Richtmyer and Morton, 1967, p. 198).

$$(T_s)_l = E_l (T_s)_{l+1} + F_l, \quad (29)$$

where $(T_s)_l$ is the soil temperature at the l th grid level from the surface. Expressions for E and F when $l > 1$ are determined from

the finite-difference form of (16), and (28) determines E_1 and F_1 .

From (28) and (29), we obtain

$$E_1 = (k_s / \Delta z_s) / [4\epsilon\sigma(T_G^n)^3 + m(P_o/P_G)^{R/c_p} + k_s / \Delta z_s] , \quad (30)$$

and

$$F_1 = [R_s + \epsilon R_{L\downarrow} + 3\epsilon\sigma(T_G^n)^4 + m\theta^n(z_1)] / [4\epsilon\sigma(T_G^n)^3 + m(P_o/P_G)^{R/c_p} + (k_s / \Delta z_s)] . \quad (31)$$

Numerical integration of (16) by use of (29) to (31) is rapid, since no iteration is required.

Finally, the longwave incoming radiation at the surface is parameterized according to Swinbank's empirical formula (quoted in Paltridge and Platt, 1976, p. 139).

$$R_{L\downarrow} = 5.31 \times 10^{-13} T^6 \cos\alpha, \quad (32)$$

where $R_{L\downarrow}$ is in $W m^{-2}$, T is the screen temperature and α , the angle of inclination of a sloped surface, is obtained from

$$\alpha = \tan^{-1} [(\partial z_g / \partial x)^2 + (\partial z_g / \partial y)^2]^{1/2}. \quad (33)$$

Numerical values of the parameters ϵ , k_s , etc. used in the actual simulations are given following a definition of each parameter in Nomenclature.

5. Initial values

As is often done in numerical simulations to avoid computational shock, the ground is initially assumed to be flat and is modified gradually with time. Accordingly, horizontally homogeneous fields

are initially assumed for all the variables to be integrated. The initial wind, and water vapor profiles (Fig. 3) are constructed from the profiles observed at PDC (Fig. 4) at midnight on 22 July, 1979. The maximum height of measurements by tethered sondes was 525 m above the ground. Thus, the profiles above this level are extrapolated from the values at lower heights. The mixing ratios in the layers above 1000 m are assumed to decrease linearly with heights and approach a constant value. Relative humidity near the ground is assumed to be 50%, consistent with the value observed. The temperatures in the soil layers are initially assumed to be constant (302 K). The potential temperatures in the atmosphere increase gradually with height from the ground values (302 K) to 305 K at height of 231 m, and above that level the potential temperatures remain constant. All the turbulence variables are initialized by using Level 2 model of Mellor and Yamada (1974). The Level 2 model is an algebraic model obtained by neglecting time dependent, advection and diffusion terms in the turbulence energy equation given by (14).

6. Numerical procedures

Following coordinate transformation by Eq. (1), procedures of numerical integration become similar to those for flow over flat terrain (Yamada, 1978a). Therefore, only a brief description of the numerical procedures is given here. The prognostic equations (2), (3) and (12) to (15) are solved by using an Alternating

Direction Implicit (ADI) method (Peaceman and Rachford, 1955), where integrations are performed in one direction at a time (Yamada, 1978a). The ADI method is numerically stable according to a linear stability analysis, but the time increment is chosen to satisfy the Courant-Friedrichs-Lewy criteria.

In order to increase the accuracy of finite-difference approximations and to suppress computational noise, mean and turbulence variables are defined on a staggered grid, both horizontally and vertically (see Fig. 2 in Yamada, 1978a). Mean wind, temperature and water vapor vary greatly with height near the surface, but much less so away from the surface. In order to resolve these variations, vertical grid levels are determined according to a log-plus-linear relationship between the grid level and the vertical coordinate (Table 1). A grid of $27 \times 20 \times 9$ points is used to cover a computational region of approximately $10 \text{ km} \times 7 \text{ km} \times 1.3 \text{ km}$. The locations of the vertical grid levels for the soil layers are also determined according to a log-plus-linear relationship (Table 1).

The vertical velocity W^* is obtained by integrating the continuity equation (4) from the ground upward. A no-slip boundary conditions, $W^* = 0$ at the ground, is used. The vertical wind is assumed to vanish along the top boundary. If W^* is not zero, minor adjustments are applied to the horizontal wind components U and V in order to satisfy the boundary condition (Yamada, 1978a).

Unlike the Level 2 model of Mellor and Yamada (1974) the present model does not cut off turbulence even when the flux Richardson number exceeds a critical value of approximately 0.21. However, eddy viscosity coefficients become very small when the flux Richardson number exceeds the critical value. A lower limit of $10^{-4} \text{ m}^2 \text{ s}^{-1}$ is imposed if eddy viscosity coefficients become smaller than the lower limit.

In this model, the lateral boundaries in south and west are designated inflow boundaries and the north and east boundaries are outflow boundaries. In other words, only the cases when inflow is between southerly and westerly are simulated. Since observations are not available at the inflow boundaries, boundary values are generated by numerical simulation of a one-dimensional (vertical) version of the governing equations (2), (3) and (12) to (15). The values at the outflow boundaries are obtained by linear extrapolation from the values one and two grid points inside from the boundary.

Variables U , V , Q_w , Θ_ℓ , q^2 , and q^2_ℓ are smoothed at each time step by using the values at four neighboring points, i.e.,

$$\phi_{i,j} = \lambda \phi_{i,j} + 0.25 \lambda (\phi_{i+1,j} + \phi_{i,j+1} + \phi_{i-1,j} + \phi_{i,j-1}),$$

where ϕ represents one of the variables and $\lambda = 0.5$ is used.

7. Results and discussions

Integration is initiated at 0000 PST and continued for 3 hours. During this period simulated wind fields nearly obtain a steady

state and the terrain variations in the model attain 99.5% of the values shown in Fig. 2. Since a staggered grid system (Section 6) is used, all variables are not computed at the same points. However, for convenience in presenting the results, variables are interpolated to common grids. Figure 4 shows the distribution of horizontal wind vectors on a terrain-following surface, 68 m above the ground. A grid point is located at the center of each arrow. The length of an arrow is proportional to the wind speed: an arrow of one grid length indicates 10 m s^{-1} . A simulation with a westerly geostrophic clearly shows downslope flow converging into a valley (Fig. 4). Drainage winds simulated along the slopes in the northwest are much larger than those in the southwest, primarily due to the difference in steepness. Horizontal winds at 68 m above the slopes in the northwest and south depart as much as 45 degrees from the generally westerly winds at 231 m (Fig. 5). Observed horizontal winds at 10 m above the ground are shown in Fig. 6. The simulated surface winds (Fig. 4) qualitatively agree with those observed (Fig. 6). However, the observed wind speeds are much smaller except at LAR (Fig. 6), where agreement is reasonably good. The primary cause for the overprediction of surface wind speeds is attributed to the irregular coverage of trees of typical height 20 m. Thus, drainage flow may have been decreased considerably by the drag due to trees. Although the immediate vicinities of observation towers are relatively clear, most observation sites are

surrounded by trees. A minimum fetch needed before the flow becomes free from the effect of upstream trees is at least 100 times the tree height. The station at LAR most nearly approaches satisfaction of the fetch requirement. This might explain why the highest wind speeds are measured at this station.

Winds simulated in the terrain-following coordinate system are interpolated at 200 m, 400 m, and 550 m above the lowest elevation (308 m msl) (Figs. 7, 8, and 9, respectively). The veering of wind with height is clearly indicated. For example, Fig. 8 shows that wind direction near the ground approximate the slope of terrain, but change rather rapidly with height to approach the westerly geostrophic wind.

Acceleration (deceleration) and convergence (divergence) of the horizontal wind vectors result in vertical winds, which can be computed from the mass conservation equation. Distribution of the vertical wind component at 145 m above the ground is shown in Fig. 10. Maximum downward motion is found over the downslope, while maximum upward motion is computed at the convergence zone over the valley. It is cautioned that the vertical motion presented here should be interpreted only qualitatively, as there has been no verification by observation. The magnitude of simulated vertical motion is a function of the height of the upper computational boundary, since vertical motion is assumed to vanish there (see Section 6). Measurements of vertical motion are urgently needed,

since vertical motion and horizontal meandering are particularly important for realistic simulation of pollutant transport over the Geysers area.

Vertical profiles of the horizontal wind speed in the cross-section through the line C-C in Fig. 11 are shown in Fig. 12. Profiles near the ground are not precise, due to the rather coarse grid resolution in the vertical direction. Nevertheless, these profiles clearly indicate wind maxima in the surface layer and weak minima in the layer above. Development of drainage flow along the slope is also evident. The upstream wind profile at the far left is almost uniform with height, but wind maxima typical to drainage flow become evident as flow travels downslope. These profiles indicate that the nocturnal drainage flow layer is very thin. According to analytical solutions for one-dimensional slope flow (Sutton, 1953), wind maxima occur at $\pi l/4$ above the ground where

$$l^4 = 4 (K_M K_H \langle \frac{H}{V} \rangle / \gamma g) \operatorname{cosec}^2 \delta .$$

In this expression K_M and K_H are eddy viscosity and eddy diffusivity, respectively; γ is a lapse rate of the potential temperature profile in the surface inversion layer, and δ is the ground slope. Substituting $K_M = K_H = 2 \text{ m}^2 \text{ s}^{-1}$ and $\gamma = 0.04 \text{ K m}^{-1}$ obtained from the simulation, and a ground slope $\delta = 10$ degrees into the above expression, one obtains $l = 25 \text{ m}$. Thus, wind maxima are expected to occur at 23 m above the ground. The wind speed profiles

(Fig. 13) in the vertical plane through the line A-A of Fig. 11 indicate wind maxima approximately 50 m above the ground. Simulated speeds of 8 m s^{-1} at 10 m above the ground agree with the observation at LAR. Figure 14 shows the wind speed profiles in the vertical cross-section through the line B-B (Fig. 11). For example, a wind speed vertical profile measured by tethered sondes at station TH7 is shown in Fig. 15. Large differences between observed and simulated wind speed profiles are noted. Computed profiles (Fig. 14) exhibit maxima near the surface, typical for drainage flow, while the observed profiles are relatively uniform with height. It is not immediately clear the reason why the observations do not exhibit the maxima predicted by theory for one-dimensional slope flow and numerical simulations. This may be due to inadequate vertical resolution in observed data, or to physical effects not considered, such as canopy drag. One way to remove the difficulty, i.e., a direct computation of tree drag, might be to simulate only the flow away from the canopy using, as surface boundary conditions, log-linear profiles for wind, temperature and water vapor with roughness lengths and displacement heights taking the tree heights into consideration. In order to know the profiles of the corresponding variables within the canopy, this model would require empirically determined similarity profiles. However, such similarity profiles within a canopy and over a sloping surface have not been developed yet.

Temperature distributions at the surface and 230 m above the ground are shown in Figs. 16 and 17, respectively. In general, simulated temperatures over the downslope surfaces are higher than those over a valley. The distribution of temperatures is correlated well with that of the simulated vertical wind (Fig. 10): when the atmosphere is stably stratified downward motion transports potentially warm air, resulting in the area of higher potential temperature and upward motion transports potentially cool air, resulting in the area of minimum potential temperature. Thus, the vertical motion, as well as turbulent and radiational heat transfers, plays an important role in determining the potential temperature profiles. Figure 18 shows profiles of the potential temperatures in a vertical plane through B-B in Fig. 11. The simulated profile of the potential temperature at PDC (Fig. 18a) agrees well with observed (Fig. 19). This agreement, however, may be fictitious because the effects of the surface tree canopy were not considered in the model. For a dense canopy, maximum cooling due to longwave radiation occurs somewhere near the crown height of trees rather than at the ground, resulting in a stable boundary layer above a canopy, and a slightly unstable or near neutral layer within a canopy. Thus, a layer of main drainage flow could be displaced above a tree canopy, which may explain an elevated wind maximum at approximately 100 m above the ground (Fig. 15). Therefore, it is very important to consider the effects of a tree canopy into the model in order to simulate

realistically both wind and temperature profiles in the Geysers area. Continuous efforts have been made to include into the model both aerodynamic and radiation effects of a tree. However, this is beyond the scope of the present study.

8. Summary and conclusions

Nocturnal drainage flow over the California Geysers area has been simulated with a three-dimensional, simplified second-moment turbulence-closure model. The simulation of nocturnal drainage flow, horizontal convergence in the valley and resulting vertical motions appears to be reasonable. Wind speed profiles exhibit increasingly large maxima in the surface layer as the flow travels downslope. Large maxima and weak minima in the simulated wind speed profiles are consistent with theoretical predictions for one-dimensional slope flow. On the other hand, observed speeds are much smaller and more uniform with height. It is quite possible that the measurements of wind profiles by tethered sondes might have missed or over-smoothed wind maxima confined to a thin layer ($<50\text{m}$) near the surface, due to a rather coarse vertical resolution. It is also speculated that the observed drainage flow may have been decreased considerably by drag induced by trees surrounding most observational sites. This speculation is partially supported by recent measurements of drainage flow in Australia over a slope with sparse tree coverage (Manins and Sawford, 1979). Their wind profiles clearly show large maxima similar to the ones predicted here. This appears to suggest that an increase of the vertical resolution in the model would not improve the agreement between computed results and observations in the Geysers area. The effects of tall

trees (canopy flow) should be taken into account, through further development of numerical models such as the present one.

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APPENDIX

Equations of motion in the transformed coordinate system

The hydrostatic equilibrium using the Boussinesq approximation may be given as (e.g., Yamada, 1978a)

$$0 = - \langle \rho \rangle^{-1} \partial P / \partial z - [1 - \beta (\bar{\theta}_v - \langle \bar{\theta}_v \rangle)] g, \quad (A1)$$

where ρ is the air density, P is the pressure, $\bar{\theta}_v$ is the virtual potential temperature and β is the thermal expansion coefficient, $(1/\langle \bar{\theta}_v \rangle)$. The bracket $\langle \rangle$ indicates an average taken over a horizontal plane. It is found convenient to eliminate the density from the equation of motion by using the equation of state.

First let us define a nondimensional pressure,

$$\Pi \equiv C_p (P/P_o)^k \quad (A2)$$

where P_o is a reference pressure (1000 mb). Differentiating (A2) logarithmically by x_i , we obtain after rearrangement

$$\partial P / \partial x_i = (P/k\Pi) \cdot \partial \Pi / \partial x_i. \quad (A3)$$

Substituting into (A3) the virtual potential temperature

$$\bar{\theta}_v = (P_o/P)^k T_v = C_p \Pi^{-1} T_v, \quad (A4)$$

and dividing by $\langle \rho \rangle$ both sides of the resultant equation, one obtains

$$\langle \rho \rangle^{-1} \partial P / \partial x_i = \rho \langle \rho \rangle^{-1} \bar{\theta}_v \cdot \partial \Pi / \partial x_i \quad (A5)$$

where the equation of state $P = \rho R_d T_v$ and a relationship $\kappa C_p = R_d$ are used. A ratio $\rho \cdot \langle \rho \rangle^{-1}$ may be rewritten as

$$\begin{aligned} \rho \cdot \langle \rho \rangle^{-1} &= 1 + \Delta \rho \cdot \langle \rho \rangle^{-1} = 1 - \Delta \bar{H}_V \cdot \langle \bar{H}_V \rangle^{-1} \\ &= 1 - \beta (\bar{H}_V - \langle \bar{H}_V \rangle) . \end{aligned} \quad (A6)$$

Substituting (A6) into (A5), we obtain

$$\langle \rho \rangle^{-1} \partial P / \partial x_i = [1 - \beta (\bar{H}_V - \langle \bar{H}_V \rangle)] \bar{H}_V \partial \Pi / \partial x_i . \quad (A7)$$

Thus, Eq. (A1) can be rewritten as

$$\partial \Pi / \partial z = -g / \bar{H}_V . \quad (A8)$$

where the term $\langle \rho \rangle^{-1} \partial P / \partial z$ is eliminated by using (A7) for $x_3 = z$.

The hydrostatic equation in the transformed coordinate system may be obtained by performing a chain differentiation of (A8), resulting

$$\partial \Pi / \partial z^* = (\partial \Pi / \partial z) \cdot (\partial z / \partial z^*) = -g (H - z_g) (\bar{H} \bar{H}_V)^{-1} , \quad (A9)$$

where Eqs. (1c) and (A8) are used.

The pressure gradient in the x direction is obtained from (A7)

$$\begin{aligned} \langle \rho \rangle^{-1} \partial P / \partial x &= [1 - \beta (\bar{H}_V - \langle \bar{H}_V \rangle)] \bar{H}_V \partial \Pi / \partial x \\ &\approx \langle \bar{H}_V \rangle \partial \Pi / \partial x , \end{aligned} \quad (A10)$$

where an approximation

$$\begin{aligned} [1 - \beta (\bar{H}_V - \langle \bar{H}_V \rangle)] \bar{H}_V &= \langle \bar{H}_V \rangle (1 - \Delta \bar{H}_V \langle \bar{H}_V \rangle^{-1}) (1 + \Delta \bar{H}_V \langle \bar{H}_V \rangle^{-1}) \\ &= \langle \bar{H}_V \rangle [1 - (\Delta \bar{H}_V / \langle \bar{H}_V \rangle)^2] \\ &\approx \langle \bar{H}_V \rangle , \end{aligned} \quad (A11a)$$

is used, Equation (A11a) is good approximation since under normal conditions

$$(\Delta \bar{H}_V / \langle \bar{H}_V \rangle)^2 \ll 1, \quad (A11b)$$

where

$$\Delta \bar{H}_V \equiv \bar{H}_V - \langle \bar{H}_V \rangle. \quad (A11c)$$

It is noted that the approximation (A11a) is consistent with the accuracy of the Boussinesq approximation used in the equations of motion. The RHS of (A10) may be rewritten in the transformed coordinate system as

$$\begin{aligned} \langle \bar{H}_V \rangle \partial \Pi / \partial x &= \langle \bar{H}_V \rangle \partial \Pi / \partial x + \langle \bar{H}_V \rangle \partial \Pi / \partial z^* \cdot \partial z^* / \partial x \\ &= \langle \bar{H}_V \rangle \partial \Pi / \partial x - g \langle \bar{H}_V \rangle (H - z_g) (\bar{H} \bar{H}_V)^{-1} \cdot \\ &\quad [(z^* - \bar{H})(H - z_g)^{-1} \cdot \partial z_g / \partial x - z^* (H - z_g)^{-1} \cdot \partial H / \partial x] \\ &= \langle \bar{H}_V \rangle \partial \Pi / \partial x - g \langle \bar{H}_V \rangle (z^* - \bar{H}) (\bar{H} \bar{H}_V)^{-1} \cdot \partial z_g / \partial x \\ &\quad + \langle \bar{H}_V \rangle \bar{H}_V^{-1} \cdot g \cdot (z^* / \bar{H}) \cdot \partial H / \partial x, \end{aligned} \quad (A12)$$

where (A9) and the following relationship obtained by differentiating (1) by x , i.e.,

$$\partial z^* / \partial x = (z^* - \bar{H})(H - z_g)^{-1} \cdot \partial z_g / \partial x - z^* (H - z_g)^{-1} \cdot \partial H / \partial x \quad (A13)$$

are used. Thus, the equation of motion in the x^* direction is

$$\begin{aligned} DU/Dt - fV &= - \langle \bar{H}_V \rangle \partial \Pi / \partial x + g \langle \bar{H}_V \rangle (z^* - \bar{H}) (\bar{H} \bar{H}_V)^{-1} \partial z_g / \partial x \\ &\quad - g \langle \bar{H}_V \rangle z^* (\bar{H} \bar{H}_V)^{-1} \partial H / \partial x + \dots \end{aligned} \quad (A14)$$

where the turbulent diffusion terms are omitted until the final expression (A25) is obtained. The pressure Π in (A14) can be

obtained by integrating (A9) from z^* to $\bar{H}$, i.e.,

$$\Pi(x, y, z^*, t) = \Pi(x, y, \bar{H}, t) + (H - z_g)(\bar{H})^{-1} \cdot g \int_{z^*}^{\bar{H}} \bar{\Theta}_V^{-1} dz'. \quad (A15)$$

Differentiation of (A15) by x yields,

$$\begin{aligned} \partial \Pi / \partial x = & \partial \Pi(\bar{H}) / \partial x + g(H - z_g)(\bar{H})^{-1} \partial \left(\int_{z^*}^{\bar{H}} \bar{\Theta}_V^{-1} dz' \right) / \partial x \\ & + g(\partial H / \partial x - \partial z_g / \partial x)(\bar{H})^{-1} \int_{z^*}^{\bar{H}} \bar{\Theta}_V^{-1} dz'. \end{aligned} \quad (A16)$$

where an abbreviated symbol $\Pi(\bar{H}) = \Pi(x, y, \bar{H}, t)$ is used. The integration in the second term on the RHS of (A16) may be rewritten as

$$\partial \left(\int_{z^*}^{\bar{H}} \bar{\Theta}_V^{-1} dz' \right) / \partial x = \int_{z^*}^{\bar{H}} (\partial \bar{\Theta}_V^{-1} / \partial x) dz' - \bar{\Theta}_V^{-1} \partial z^* / \partial x, \quad (A17)$$

where the Leibnitz rule of exchanging integration and differentiation, and $\partial \bar{H} / \partial x = 0$ are used. Deviations of $\bar{\Theta}_V$ from $\langle \bar{\Theta}_V \rangle$ are small under normal conditions, i.e., $\Delta \bar{\Theta}_V / \langle \bar{\Theta}_V \rangle \ll 1$, so that following approximation holds:

$$\bar{\Theta}_V^{-1} \approx \langle \bar{\Theta}_V \rangle^{-1} (1 - \Delta \bar{\Theta}_V \cdot \langle \bar{\Theta}_V \rangle^{-1}). \quad (A18)$$

where $\Delta \bar{\Theta}_V$ is already defined by (A11c). Differentiation of (A18) by x yields

$$\partial \bar{\Theta}_V^{-1} / \partial x \approx - \langle \bar{\Theta}_V \rangle^{-2} \cdot \partial \Delta \bar{\Theta}_V / \partial x. \quad (A19)$$

Substituting (A13) and (A19) into (A17), we obtain

$$\begin{aligned} \partial \left(\int_{z^*}^{\bar{H}} \bar{\Theta}_V^{-1} dz' \right) / \partial x = & - \int_{z^*}^{\bar{H}} \langle \bar{\Theta}_V \rangle^{-2} \cdot \partial \Delta \bar{\Theta}_V / \partial x \cdot dz' \\ & - \bar{\Theta}_V^{-1} (H - z_g)^{-1} [(z^* - \bar{H}) \cdot \partial z_g / \partial x - z^* \cdot \partial H / \partial x]. \end{aligned} \quad (A20)$$

The integration in the third term on the RHS of (A16) is also approximated by using a relationship (A18), resulting

$$\int_{z^*}^{\bar{H}} \bar{\Theta}_v^{-1} dz' \approx \langle \bar{\Theta}_v \rangle^{-1} [(\bar{H} - z^*) - \int_{z^*}^{\bar{H}} \Delta \bar{\Theta}_v dz'] \quad (A21)$$

where $\langle \bar{\Theta}_v \rangle$ is assumed to be independent with height. This assumption is not in general true. However, the errors which might result from using the approximation (A21) are probably not very significant.

Substituting (A20) and (A21) into (A16), we obtain after rearrangement

$$\begin{aligned} \partial \Pi / \partial x = & \partial \Pi(\bar{H}) / \partial x + g(\bar{H} - z^*)(\bar{H})^{-1} [\bar{\Theta}_v^{-1} - \langle \bar{\Theta}_v \rangle^{-1}] \partial z_g / \partial x \\ & + g[\bar{\Theta}_v^{-1} (\bar{H})^{-1} z^* + \langle \bar{\Theta}_v \rangle^{-1} (\bar{H})^{-1} (\bar{H} - z^*)] \partial H / \partial x \\ & - g(\bar{H} - z_g)(\bar{H})^{-1} \int_{z^*}^{\bar{H}} \langle \bar{\Theta}_v \rangle^{-2} \cdot \partial \Delta \bar{\Theta}_v / \partial x \cdot dz' \\ & - g \langle \bar{\Theta}_v \rangle^{-1} (\bar{H})^{-1} (\partial H / \partial x - \partial z_g / \partial x) \int_{z^*}^{\bar{H}} \Delta \bar{\Theta}_v dz'. \end{aligned} \quad (A22)$$

Substituting (A22) into the equation of motion (A14), we obtain

$$\begin{aligned} DU/Dt - fV = & - \langle \bar{\Theta}_v \rangle \partial \Pi(\bar{H}) / \partial x \\ & - g(\bar{H} - z^*)(\bar{H})^{-1} [2 \langle \bar{\Theta}_v \rangle \bar{H}_v^{-1} - 1] \partial z_g / \partial x \\ & - g[2 \langle \bar{\Theta}_v \rangle z^* \bar{\Theta}_v^{-1} (\bar{H})^{-1} - (\bar{H} - z^*)(\bar{H})^{-1}] \cdot \partial H / \partial x \\ & + g(\bar{H} - z_g)(\bar{H})^{-1} \int_{z^*}^{\bar{H}} \langle \bar{\Theta}_v \rangle^{-1} \cdot \partial \Delta \bar{\Theta}_v / \partial x \cdot dz' \\ & + g(\bar{H})^{-1} (\partial H / \partial x - \partial z_g / \partial x) \int_{z^*}^{\bar{H}} \Delta \bar{\Theta}_v dz'. \end{aligned} \quad (A23)$$

The following conditions along the top boundary ($z^* = \bar{H}$) are assumed: $V = V_g(\bar{H})$, $DU/Dt = 0$ and $\langle \bar{\Theta}_v(\bar{H}) \rangle = \bar{\Theta}_v(\bar{H})$ where V_g is the y component of the geostrophic wind. Substituting these boundary conditions into (A23), we obtain after rearrangement

$$\langle \hat{\Theta}_V(\bar{H}) \rangle \partial \Pi(\bar{H}) / \partial x = fV_g(\bar{H}) - 2g \cdot \partial H / \partial x. \quad (A24)$$

Eliminating $\partial \Pi(\bar{H}) / \partial x$ in (A23) by using (A24), we finally obtain the following x component of equation of motion after rearrangement

$$\begin{aligned} DU/Dt = & f(V - V_g) + g(\bar{H} - z_g^*)(\bar{H})^{-1} [1 - 2\langle \hat{\Theta}_V \rangle \langle \hat{\Theta}_V \rangle^{-1}] \partial z_g / \partial x \\ & + g[2\langle \hat{\Theta}_V \rangle \langle \hat{\Theta}_V(\bar{H}) \rangle^{-1} - (2\langle \hat{\Theta}_V \rangle \langle \hat{\Theta}_V \rangle^{-1} - 1) z_g^*(\bar{H})^{-1} - 1] \\ & \cdot \partial H / \partial x + \partial(K_x \partial U / \partial x) / \partial x + \partial(K_{xy} \partial V / \partial y) / \partial y \\ & + \bar{H}(H - z_g)^{-1} \partial(-\overline{uw}) / \partial z^*, \end{aligned} \quad (A25)$$

where

$$\begin{aligned} fV_g \equiv & fV_g(\bar{H}) \cdot \langle \hat{\Theta}_V \rangle \cdot \langle \hat{\Theta}_V(\bar{H}) \rangle^{-1} \\ & - g(H - z_g)(\bar{H})^{-1} \int_{z_g^*}^{\bar{H}} \langle \hat{\Theta}_V \rangle^{-1} \cdot \partial \Delta \hat{\Theta}_V / \partial x \cdot dz' \\ & - g(\bar{H})^{-1} (\partial H / \partial x - \partial z_g / \partial x) \int_{z_g^*}^{\bar{H}} \Delta \hat{\Theta}_V dz', \end{aligned} \quad (A26)$$

In a similar fashion, the y component of equation of motion is obtained (Eq. 3).

Boundary conditions

The boundary conditions in the surface layer are obtained from the following formulas:

$$\frac{\tilde{V}(z)}{u_*} = \frac{1}{k} \left[\ln \left\{ (z + z_o) / z_o \right\} - \Psi_m(\zeta) \right], \quad (A27)$$

$$\frac{\Theta_l(z) - \Theta_l(o)}{T_*} = \frac{P_r}{k} \left[\ln \left\{ (z + z_{ot}) / z_{ot} \right\} - \Psi_h(\zeta) \right], \quad (A28)$$

$$\frac{Q_w(z) - Q_w(o)}{Q_*} = \frac{S_c}{k} \left[\ln \left\{ (z + z_{ov})/z_{ov} \right\} - \Psi_v(\zeta) \right] \quad , \quad (A29)$$

$$q^2(z) = B_1^{2/3} u_*^2 (\Phi_m - \zeta)^{2/3} \quad , \quad (A30)$$

and

$$q^2_l(z) = kz q^2(z) \quad (A31)$$

In the above equations, $\tilde{V}(z)$, $\Theta_l(z)$, and $Q_w(z)$ are the abbreviations for $\tilde{V}(x, y, z, t)$, $\Theta_l(x, y, z, t)$, and $Q_w(x, y, z, t)$, respectively and $\tilde{V}$ is the horizontal wind speed $(U^2 + V^2)^{1/2}$. Terms u_* , T_* , and Q_* , defined as $u_* \equiv \sqrt{\tau/\rho}$, $T_* \equiv H/\rho c_p u_*$, and $Q_* \equiv E/\rho u_*$, are friction velocity and scales for temperature and water vapor, respectively. Terms τ , H , and E are surface stress, total sensible heat, and rate of evaporation, respectively; total sensible heat being defined as $H = \overline{w\theta}_v = (1 + 0.61 Q_w) \overline{w\theta} + 0.61 \overline{wq}_w$. The parameter ζ is a nondimensional height z/L , where L is the Obukhov length $-u_*^3/k\beta gH$; k is the von Karman constant; z_o , z_{ot} , and z_{ov} are roughness lengths for wind, temperature, and water vapor, respectively; and P_r and S_c are the turbulence Prandtl and Schmidt numbers, respectively. Terms Ψ_m , Ψ_h , and Ψ_v (Panofsky, 1963) are correction terms for the atmospheric stability and their functional forms are given by

$$\Psi_m(\zeta) = \int_0^\zeta (1 - \Phi_m(\zeta'))/\zeta' d\zeta' \quad , \quad (A32)$$

$$\Psi_h(\zeta) = \int_0^\zeta (1 - \Phi_h(\zeta'))/\zeta' d\zeta' \quad , \quad (A33)$$

and

$$\Psi_v(\zeta) = \int_0^\zeta (1 - \Phi_v(\zeta'))/\zeta' d\zeta' \quad , \quad (A34)$$

where Φ_m , Φ_h , and Φ_v are nondimensional wind, temperature, and water vapor gradients, respectively. The following formulations (Dyer and Hicks, 1970) are used for Φ_m , Φ_h , and Φ_v under unstable conditions,

$$\Phi_m(\zeta) \equiv (u_* / kz) (\partial U / \partial z) = (1 - 15 \zeta)^{-1/4} , \quad (\text{A35})$$

$$\Phi_h(\zeta) \equiv (T_* / kz) (\partial \Theta_f / \partial z) = (1 - 15 \zeta)^{-1/2} , \quad (\text{A36})$$

and

$$\Phi_v(\zeta) \equiv (Q_* / kz) (\partial Q_w / \partial z) = (1 - 15 \zeta)^{-1/2} . \quad (\text{A37})$$

For stable conditions

$$\Phi_m(\zeta) = \Phi_h(\zeta) = \Phi_v(\zeta) = 1 + 5 \zeta . \quad (\text{A38a,b,c})$$

The roughness lengths z_o , z_{ot} , and z_{ov} are specified over land.

The boundary conditions at the top are assumed to be

$$\partial U / \partial z = \partial V / \partial z = 0 , \quad (\text{A39a,b})$$

$$\partial \Theta_f / \partial z = 0 , \quad (\text{A40})$$

$$\partial Q_w / \partial z = 0 , \quad (\text{A41})$$

$$q^2 = 0 , \quad (\text{A42})$$

and

$$\partial (q^2 f) / \partial z = 0 . \quad (\text{A43})$$

LIST OF SYMBOLS AND NUMERICAL VALUES USED

C_p	specific heat capacity of dry air at constant pressure, 1.003 joules $g^{-1} K^{-1}$
E_1, E_2	empirical constants in a length scale equation (15), $E_1 = 1.8$ and $E_2 = 1.33$ from Mellor and Yamada (1977)
k	Von Karman constant, 0.4
k_s	soil diffusivity, $1.11 \times 10^{-6} m^2 s^{-1}$
L	latent heat of water vapor, 2500 joules g^{-1}
l	a master turbulence length scale
P	pressure
P_o	a reference pressure, 1000 mb
P_r	turbulent Prandtl number for neutral condition, 1.0
q^2	twice the turbulence energy, $\overline{\sum u_i u_i}$
q_w	fluctuation of mixing ratio of total water
Q_v, Q_ℓ, Q_w	mean mixing ratios of water vapor, liquid water and total water
R	gas constant for dry air, 0.28704 joules $g^{-1} K^{-1}$
u_*	friction velocity
$\overline{u_k q_w}$	turbulent moisture flux
$\overline{u_k u_j}$	Reynolds stress
$\overline{u_k \theta_\ell}$	turbulent heat flux
β	thermal expansion coefficient, $\approx 1/\bar{\theta}_v$
ϵ	emissivity, 0.98
θ_v	fluctuation of virtual potential temperature
$\bar{H}_\ell$	liquid potential temperature, $\equiv (\bar{\theta} - \bar{\theta}_T) (L/c_p) Q_\ell$
$\bar{H}_v$	virtual potential temperature, $\equiv \bar{\theta}(1 + 0.61Q_w - 1.61Q_\ell)$

k ratio R/c_p , 0.286

σ Stefan-Boltzman constant, $5.67 \times 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$

— ensemble average

	Level #	Level (m)	Intermediate level (m)
Atmosphere	9	1280	1182
	8	1085	988
	7	891	794
	6	698	602
	5	508	414
	4	322	231
	3	145	68
	2	15	0.8
	1	0	0.003
	2	0.009	0.024
Soil	3	0.054	0.100
	4	0.159	0.228
	5	0.303	0.383
	6	0.466	0.551
	7	0.638	0.727
	8	0.817	0.907
	9	1.0	

Table 1. Locations of the vertical grid levels used for atmospheric and soil layers.

Figure 1

Topography in the Geysers area. Simulations are performed over the area enclosed.

Figure 2

Three-dimensional presentation of the area simulated.

Figure 3

Initial profiles of a) wind and b) mixing ratio of water vapor.

Figure 4

Computed wind vectors at 68 m above the ground. Symbols in the figure indicate observational sites. Terrain is contoured by solid lines with an increment of 100 m; the lowest contour is at 400 m above the mean sea level.

Figure 5

Same as in Fig. 4 except at 231 m above the ground.

Figure 6

Observed wind vectors at 10 m above the ground, averaged over the entire experimental period. The wind vector at LAR represents approximately 10 m s^{-1} .

Figure 7

Computed horizontal wind vectors 200 m above the lowest point in the area. Terrain is contoured by solid lines with an increment of 100 m; the lowest contour is at 400 m above the mean sea level.

Figure 8

Same as in Fig. 7 except at 400 m above the lowest point.

Figure 9

Same as in Fig. 7 except at 550 m above the lowest point.

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Figure 10
 Computed vertical velocities (cm s^{-1}) at 145 m above the ground.

Figure 11

A map to show the locations of the cross sections used in Figs. 12 to 14.

Figure 12

Computed wind speed profiles in the vertical plane through C-C in Fig. 11. Exact locations are given by grid coordinates (I,J).

Figure 13

Same as in Fig. 12 except in the vertical plane through A-A in Fig. 11.

Figure 14

Same as in Fig. 12 except in the vertical plane through B-B in Fig. 11.

Figure 15

Observed wind speed profile at PDC at
0400 PST.

Computed potential temperature (K) at the surface.
 Symbols in the figure and terrain contours are the
 same as in Fig. 4.

Figure 16

Figure 17

Same as in Fig. 16 except at 231 m above the ground.

Figure 18

Computed potential temperature profiles in a vertical plane through B-B in Fig. 11. Exact locations are given by grid coordinates (I,J). a) (3,17), b) (11,14), and c) (19,10).

Figure 19

Observed potential temperature profile at PDC at 0400 PST.

AN ANALYTICAL SOLUTION FOR THREE-DIMENSIONAL STATIONARY FLOWS
IN THE ATMOSPHERIC BOUNDARY LAYER OVER TERRAIN

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ABSTRACT

An analytical solution to the Navier-Stokes equations of motion for three-dimensional stationary flow in the atmospheric boundary layer over terrain is presented. Analyses of the effects of topography and the Coriolis force on the velocity distribution in the boundary layer indicate that (1) upward motion occurs on the windward side and the slope to the right (left) of the geostrophic wind in the northern (southern) hemisphere, whereas downward motion occurs on the lee side and the slope to the left (right) of the geostrophic wind in the northern (southern) hemisphere, and (2) the horizontal component of the velocity turns right (left) in the northern (southern) hemisphere with increasing height.

1. INTRODUCTION

Velocity distribution in the atmospheric boundary layer over a flat surface was first studied by Ekman (1905) who assumed a constant eddy viscosity in the planetary boundary layer, and obtained an exact solution to the Navier-Stokes equations for the balance between Coriolis, pressure-gradient, and viscous forces.

In recent years, a large number of analytical and numerical models for the study of the mean and turbulent motions in the planetary boundary layer under various thermal stratifications has been constructed. However, most of these investigations have emphasized on flows in the boundary layer over flat surfaces.

Because of increasing concern about atmospheric pollution in many population centers, industrial and power plants, which are located in valleys and terrain, and since atmospheric motion is the mechanism for the transport and dispersion of pollutants, there is a growing interest in the atmospheric motion and pollution in these regions (Kao, 1976).

For flows in the planetary boundary layer over flat surfaces, the mean motion may be assumed to be homogeneous in the horizontal, therefore, the Navier-Stokes equations become linear (Ekman, 1905) and the motion is horizontal. However, when the inhomogeneity in the topographical configuration of the earth's surface is taken into account, the motion is three-dimensional and the equations of motion are no longer linear. The purposes of this note are to seek an analytical solution to the Navier-Stokes equations for three-dimensional stationary flows in the atmospheric boundary layer over mountain-terrain, and to analyze the effects of topography and Coriolis force on the velocity distribution in the boundary layer.

2. AN ANALYTIC SOLUTION TO THE BOUNDARY LAYER EQUATIONS

We consider a stationary flow in the planetary and surface boundary layers over mountain terrain. Let $h(x,y)$ be the topographical configuration of the terrain, h_s be the surface boundary thickness, and $G = u_g + i v_g$ is the geostrophic wind velocity.

For the stationary flow in the planetary boundary layer, the Navier-Stokes equations and the equation of continuity may be express as

$$u \frac{\partial u}{\partial x} + v \frac{\partial u}{\partial y} + w \frac{\partial u}{\partial z} = f(v - v_g) + K \frac{\partial^2 u}{\partial z^2} \quad (1)$$

$$u \frac{\partial v}{\partial x} + v \frac{\partial v}{\partial y} + w \frac{\partial v}{\partial z} = -f(u - u_g) + K \frac{\partial^2 v}{\partial z^2} \quad (2)$$

$$\frac{\partial P}{\partial z} = -\rho g \quad (3)$$

$$\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y} + \frac{\partial w}{\partial z} = 0 \quad (4)$$

where u , v , w are respectively the x , y , z components of the velocity, p is the pressure, g is the gravity acceleration, ρ is the density, f is the Coriolis parameter, and K is the coefficient of eddy diffusivity.

For the stationary flow in the surface boundary layer, the equation of motion may be expressed as

$$\frac{\partial}{\partial z} |u + i v| = \frac{u_*}{k} \left(\frac{1}{z} + \frac{a}{L} \right) \quad (5)$$

where $i = (-1)^{1/2}$, $u_* = (\tau/\rho)^{1/2}$, a is a constant which has been experimentally determined to be 4.75, L is the Monin-Obukhov length, and k is the Von Karman's constant.

The boundary conditions for this model are that all components of the velocity are zero at the solid boundary, $z = h(x, y)$, the velocity tends to the geostrophic velocity as z tends to infinity, and that at the lower boundary of the planetary boundary layer the wind direction coincides with the wind stress, i.e.,

$$u = v = w = 0, \quad \text{at } z = h(x, y) \quad (6)$$

$$u \rightarrow u_g, \quad v \rightarrow v_g, \quad \text{as } z \rightarrow \infty \quad (7)$$

$$u + iv = A \frac{\partial}{\partial z} (u + iv) , \quad \text{at } z = h(x, y) + h_s \quad (8)$$

where A is a constant.

Combining Eqs. (1) and (2), we obtain

$$\left(u \frac{\partial}{\partial x} + v \frac{\partial}{\partial y} + w \frac{\partial}{\partial z} \right) (u + iv - G) = \left(K \frac{\partial^2}{\partial z^2} - if \right) (u + iv - G) . \quad (9)$$

Differential equation (9) has the following solution which satisfies the boundary conditions (7):

$$(u + iv) = G + (B_r + iB_i) e^{-(1+i)\nu[z-h(x,y)-h_s]} \quad (10)$$

$$w = \left(u_g + \exp \left\{ -\nu [z-h(x,y) - h_s] \right\} \right.$$

$$\left. \left\{ B_r \cos \nu [z-h(x,y)-h_s] + B_i \sin \nu [z-h(x,y)-h_s] \right\} \right) \frac{\partial h}{\partial x}$$

$$+ \left(v_g + \exp \left\{ -\nu [z-h(x,y) - h_s] \right\} \right.$$

$$\left. \left\{ B_i \cos \nu [z-h(x,y)-h_s] - B_r \sin \nu [z-h(x,y)-h_s] \right\} \right) \frac{\partial h}{\partial y} \quad (11)$$

where B_r and B_i are constants, and $\nu = (f/2K)^{1/2}$.

Since wind in the surface boundary layer is uni-directional, wind velocity in the surface boundary layer may be expressed as

$$u + iv = |u + iv| e^{i\alpha} \quad (12)$$

where α is the angle made between the wind in the surface boundary layer and the geostrophic wind.

At the lower boundary of the planetary boundary layer, Eq. (10) yields

$$(u + iv)_{z=h(x,y)+h_s} = G + (B_r + iB_i) \quad (13)$$

Elimination of the left-hand-side terms of Eqs. (12) and (13) gives

$$(B_r + iB_i) = |u + iv|_{z=h(x,y)+h_s} e^{i\alpha} - G \quad (14)$$

Equation (10) may then be written as

$$u + iv = G + \{ |u + iv|_{z=h(x,y)+h_s} e^{i\alpha} - G \} e^{-(1+i)v[z-h(x,y)-h_s]} \quad (15)$$

for $z \geq h(x,y)+h_s$.

It can be shown by applying the boundary condition (8) to Eq. (15) that

$$|u + iv|_{z=h(x,y)+h_s} \{ A v (\cos \alpha - \sin \alpha) + \cos \alpha \} - A v (u_g - v_g) = 0 \quad (16)$$

$$|u + iv|_{z=h(x,y)+h_s} \{ A v (\cos \alpha - \sin \alpha) + \sin \alpha \} - A v (u_g + v_g) = 0 \quad (17)$$

Solving for A and $|u + iv|_{z=h(x,y)+h_s}$ from Eqs. (16 and (17), we obtain

$$A = \frac{u_g (\cos \alpha - \sin \alpha) + v_g (\cos \alpha + \sin \alpha)}{2v(u_g \sin \alpha - v_g \cos \alpha)} \quad (18)$$

$$|u + iv|_{z=h(x,y)+h_s} = u_g (\cos \alpha - \sin \alpha) + v_g (\cos \alpha + \sin \alpha) \quad (19)$$

Substitution of (19) into (15) gives

$$u + i v = (u_g + i v_g) e^{-(1+i)v[z-h(x,y)-h_s]}$$

$$\left\{ (u_g + i v_g) - \left[u_g (\cos \alpha - \sin \alpha) + v_g (\cos \alpha + \sin \alpha) \right] e^{i\alpha} \right\} \quad (20)$$

The real and imaginary parts of Eq. (20) can be shown to be respectively,

$$u = u_g - \exp\left\{-v[z-h(x,y)-h_s]\right\} \left\{ u_g \cos v[z-h(x,y)-h_s] + v_g \sin v[z-h(x,y)-h_s] \right. \\ \left. - \left[u_g (\cos \alpha - \sin \alpha) + v_g (\cos \alpha + \sin \alpha) \right] \cos(v[z-h(x,y)-h_s] - \alpha) \right\} \quad (21)$$

$$v = v_g - \exp\left\{-v[z-h(x,y)-h_s]\right\} \left\{ -u_g \sin v[z-h(x,y)-h_s] + v_g \cos v[z-h(x,y)-h_s] \right. \\ \left. + \left[u_g (\cos \alpha - \sin \alpha) + v_g (\cos \alpha + \sin \alpha) \right] \sin(v[z-h(x,y)-h_s] - \alpha) \right\} \quad (22)$$

and

$$w = u \frac{\partial h}{\partial x} + v \frac{\partial h}{\partial y} \quad (23)$$

To match the wind distribution in the surface boundary layer with the wind at the lower boundary of the planetary layer, let the wind distribution in the surface boundary layer be

$$|u + i v| = \frac{u_*}{k} \left(\ln \frac{z-h(x,y) + z_0}{z_0} + a \frac{z-h(x,y)}{L} \right) \quad (24)$$

At the lower boundary of the planetary boundary layer, Eq. (5) becomes

$$|u + i v|_{z=h(x,y)+h_s} = \frac{u_*}{k} \left(\ln \frac{h_s+z_0}{z_0} + a \frac{h_s}{L} \right) \quad (25)$$

Elimination of the right-hand side terms from (19) and (25) gives

$$u_* = \frac{k[u_g(\cos \alpha - \sin \alpha) + v_g(\cos \alpha + \sin \alpha)]}{\left(\ln \frac{h_s + z_0}{z_0} + a \frac{h_s}{L}\right)} \quad (26)$$

It may be noticed that friction velocity, u_* , depends on G , L and .

Substitution of (26) into (24) gives the velocity profiles in the surface boundary layer,

$$|u + iv| = \frac{[u_g(\cos \alpha + \sin \alpha) + v_g(\cos \alpha - \sin \alpha)]}{\left(\ln \frac{h_s + z_0}{z_0} + a \frac{h_s}{L}\right)} \left(\ln \frac{z - h(x, y) + z_0}{z_0} + a \frac{z - h(x, y)}{L}\right) \quad (27)$$

$$w = |u + iv| \left(\cos \alpha \frac{\partial h}{\partial x} + \sin \alpha \frac{\partial h}{\partial y}\right)$$

for $h(x, y) \leq z \leq h(x, y) + h_2$.

If the x -axis is oriented parallel to the geostrophic wind, α is therefore the cross isobaric angle of the wind at the surface, and Eq. (2) becomes

$$u + iv = G + G\{(\cos \alpha - \sin \alpha) e^{i\alpha} - 1\} e^{-(1+i)\nu[z - h(x, y) - h_s]}$$

$$= G + G \sin \{\alpha - (\sin \alpha + \cos \alpha) + i(\cos \alpha - \sin \alpha)\} e^{-(1+i)\nu[z - h(x, y) - h_s]} \quad (28)$$

Since

$$e^{i(\alpha + (3/4)\pi)} = \cos(\alpha + (3/4)\pi) + i \sin(\alpha + (3/4)\pi), \quad (29)$$

Eq. (28) may be written as

$$u + iv = G \left\{ 1 + \sqrt{2} \sin \alpha e^{-\nu [z-h(x,y)-h_s]} e^{i(\alpha + (3/4)\pi - \nu [z-h(x,y)-h_s])} \right\}. \quad (30)$$

Therefore,

$$u = G \left\{ 1 + 2^{1/2} \exp \left(-\nu [z-h(x,y)-h_s] \right) \sin \alpha \cos \left(\alpha + \frac{3}{4} \pi - \nu [z-h(x,y)-h_s] \right) \right\} \quad (31)$$

$$v = 2^{1/2} G \exp \left(-\nu [z-h(x,y)-h_s] \right) \sin \alpha \sin \left(\alpha + \frac{3}{4} \pi - \nu [z-h(x,y)-h_s] \right) \quad (32)$$

$$w = G \left\{ 1 + 2^{1/2} \exp \left(-\nu [z-h(x,y)-h_s] \right) \sin \alpha \cos \left(\alpha + \frac{3}{4} \pi - \nu [z-h(x,y)-h_s] \right) \right\} \frac{\partial h}{\partial x} \\ + 2^{1/2} G \exp \left(-\nu [z-h(x,y)-h_s] \right) \sin \alpha \sin \left(\alpha + \frac{3}{4} \pi - \nu [z-h(x,y)-h_s] \right) \frac{\partial h}{\partial y} \quad (33)$$

Eq. (27) becomes

$$|u + iv| = \frac{G(\cos \alpha - \sin \alpha)}{\left(\ln \frac{h_s + z_0}{z_0} + a \frac{h_s}{L} \right)} \left(\ln \frac{z-h(x,y)+z_0}{z_0} + a \frac{z-h(x,y)}{L} \right) \quad (34)$$

$$w = |u + iv| \left(\cos \alpha \frac{\partial h}{\partial x} + \sin \alpha \frac{\partial h}{\partial y} \right)$$

for $h(x,y) \leq z \leq h(x,y) + h_s$.

In Eqs. (30), (31), (32) and (34), the cross isobaric angle α may be determined from the condition that $v = 0$ at the geostrophic wind level, $z = H$. Thus, by virtue of Eq. (32), must

$$\alpha = \left(\frac{\Omega \sin \phi}{K} \right)^{1/2} \left[H - h(x,y) - h_s \right] - \frac{3}{4} \pi \quad (35)$$

Eq. (35) indicates that the cross isobaric angle is a function of the coefficient of eddy diffusivity, topographical configuration, and latitude, ϕ , and that α decreases with increasing height of the topography.

Substitutions of (35) into Eqs. (30), (31), (32) and (34) yield respectively

$$u = G \left\{ 1 + 2^{1/2} \exp(-\nu[z-h(x,y)-h_s]) \sin(\nu[H-h(x,y)-h_s] - \frac{3}{4}\pi) \cos \nu(H-z) \right\} \quad (36)$$

$$v = 2^{1/2} G \exp(-\nu[z-h(x,y)-h_s]) \sin(\nu[H-h(x,y)-h_s] - \frac{3}{4}\pi) \sin \nu(H-z) \quad (37)$$

$$w = G \frac{\partial h}{\partial x} + 2^{1/2} G \exp(-\nu[z-h(x,y)-h_s]) \sin(\nu[H-h(x,y)-h_s] - \frac{3}{4}\pi)$$

$$\left\{ \cos \nu(H-z) \frac{\partial h}{\partial x} + \sin \nu(H-z) \frac{\partial h}{\partial y} \right\} \quad (38)$$

$$|u + iv| = \frac{2^{1/2} G \cos(\nu[H-h(x,y)-h_s] - 1/2\pi)}{\left(\ln \frac{h_s + z_0}{z_0} + a \frac{h_s}{L} \right)} \cdot \left(\ln \frac{z-h(x,y)+z_0}{z_0} + a \frac{z-h(x,y)}{L} \right) \quad (39)$$

3. FLOW CHARACTERISTICS IN THE ATMOSPHERIC BOUNDARY LAYER OVER A CIRCULAR MOUNTAIN OF SMALL HEIGHT

To analyze the effect of a mountain of small height on the velocity distribution in the planetary boundary layer, we choose a model mountain of which the height takes the following non-dimensional form

$$h^*(x^*, y^*) = a\nu H \left\{ \frac{1}{1 + e^{-b[(x^{*2} + y^{*2})^{1/2} + c]}} + \frac{1}{1 + e^{b[(x^{*2} + y^{*2})^{1/2} - c]}} - 1 \right\} \quad (40)$$

where $x^* = vx$, $y^* = vy$, $h^* = vh$, $h_s^* = vh_s$ are respectively the non-dimensionalized coordinates x , y , height of the mountain, and the surface boundary thickness.

For $a = b = c = 1$, $vH = 1$, $h^*(0,0) = 0.1$, $h_s^* = 0.1$, and $z_0^* = 0.0001$, we have computed the non-dimensionalized velocity components, $u^* = u/G$, $v^* = v/G$ and $w^* = w/G$ with the use of equations (36) through (39), and plotted the results on Figs. 1 and 2.

Figure 1 shows the distribution of the non-dimensional velocity components in the vertical cross-section passing through the mountain top, parallel to the geostrophic wind. It is seen that $u = v = w = 0$ at the surface of the mountain, and that $u \rightarrow G$, $v \rightarrow 0$ as $z \rightarrow \infty$. On the windward slope of the mountain, a horizontal convergence of u results in an upward motion, where as on the lee side of the mountain a downward motion occurs as a consequence of a horizontal divergence of u .

Figure 2 shows the distribution of the non-dimensional velocity components in the vertical cross-section perpendicular to the geostrophic wind, passing through the mountain top. It is seen that $u = v = w = 0$ at the surface of the mountain, and that $u \rightarrow G$, $v \rightarrow 0$ as $z \rightarrow \infty$. On the slope to the right of the mountain top, in the northern (southern) hemisphere, a horizontal convergence (divergence) of v results in an upward (downward) motion, whereas on the slope to the left of the mountain top, in the northern (southern) hemisphere, a horizontal divergence (convergence) of v contributes to a downward (upward) motion.

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FIGURE 1. Distribution of non-dimensional velocity components in the vertical cross section passing through the mountain top, parallel to the geostrophic wind.

FIGURE 2. Same as Fig. 1 except the vertical cross section is perpendicular to the geostrophic wind.

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APPLICATION OF DIGITAL IMAGE ANALYSIS TECHNIQUES TO
THE GEYSER'S DATA AND TOPOGRAPHY

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INTRODUCTION

This paper describes the results of digital image analysis and techniques applied to acoustic sounder data and topographic relief in the Geyser's region. Application of two dimensional digital Fourier transformation is a common technique used in image analysis.¹ First an image is transformed and then the transformation is filtered to stress high or low spacial frequency variations and an inverse transformation is applied for image enhancement. The two dimensional fast Fourier transform (2DFFT) therefore represents the spacial variability of a photographic image. The spacial variability of topography in complex terrain can be represented in this way and insight into degree of complexity and dominating spacial wavelengths can be gained. This was performed for a 16 km square digitized topographic map of the Geyser's region with 63.5 m resolution.

It was also of interest to compare facsimile recordings of acoustic sounder data to optical turbulence measurements. Since the darkness of the recording is a function of the temperature structure parameter C_T it was necessary to digitize photographic negatives of these chart recordings. C_T could then be related to the

index of refraction structure parameter C_n through the standard deviation of the log of laser light intensity σ_I measured with an optical anemometer.

RESULTS

This section is divided in two parts. The first part describes the results of a 2DFFT of Geyser's area topography. The second part describes the results of digitizing the acoustic sounder records and comparing to optical turbulence measurements.

The Fourier transform simplifies, in one respect, interpretation of complex topography by presenting data on a spacial frequency or wavelength basis. The 2DFFT, however, doesn't look like anything a person actually sees unless one looks at a transparency of intensity proportional to altitude through a lens at twice the focal length rather than in focus. The variability of the transform from corner to corner do corresponds to the axis of the original image but individual points on and off the axis do not correspond to individual points on the digitized topography. In a 2DFFT highly variable terrain tends to spread out the transform coefficients to higher values at smaller spatial wavelengths. Regularly varying terrain would cause peaks and valleys in the Fourier coefficients. These characteristics can be compared for different geographic locations and questions like the following can be answered more easily:

1. Which potential location for complex terrain studies is more complex?
2. What are the scales of regular oscillation in the terrain?
3. What is the relative importance of these regular terrain features?

To answer the first question a comparison is necessary between 2DFFT's. The latter two questions can be answered by looking at the 2DFFT generated for the

Geyser's region. The next paragraph describes the generation and interpretation of the 2DFFT.

A data file of digitized topography with 63.5 metre resolution was obtained for the area surrounding the location of the 1979 ASCOT intensive experiment. Since the array size for a 2DFFT must be a power of 2, a 256 by 256 array of altitude versus position was selected. This corresponds to an area 16.26 km on a side. A contour map of this area is shown in Fig. 1. The contours interpolated on this map were much coarser than the 63.5 m resolution actually used in the 2DFFT. Also altitude above sea level was given in 50 m increments. The highest area (Cobb Mountain) corresponds to a height of 1439 m and the lowest point was 278 m above sea level. The resultant contours of Fourier coefficients from the 2DFFT are shown in Fig. 2. These were hand drawn from a false color image of the 2DFFT. The center of the 2DFFT corresponds to a spacial wavelength of the topograph array size of 16.26 km. The corners and sizes correspond to wavenumber space representations of the two horizontal axes Fourier coefficients. The Nyquist spacial wavelength is twice the maximum resolution or 127 meters. The value of the Fourier coefficient is highest at the center of the 2DFFT and equals 256 squared times the average height above sea level (656 m). The high value at the center means that a complicated map would reproduce itself exactly at a spacial wavelength corresponding to the size of the map. What is of real importance is how the 2DFFT varies along the diagonals from corner to corner in the 2DFFT. Going from the lower right to the upper left corner of the 2DFFT corresponds to variability associated with going from southeast to northwest corner of the topography shown in Fig. 1. Inspection of the digitized topography in Fig. 1 shows that the topography varies much more regularly along the southwest to northeast diagonal than the other diagonal with about eight ridge valley systems of similar

widths. These and other features are also shown in the 2DFFT in Fig. 2. Inspection of Fig. 2 reveals the following:

1. Along the diagonal from lower right to upper left (southeast to northwest) there is much more high spacial frequency variability than along the other diagonal. The presence of many higher frequency bands above once per 8 km may be due to stream erosion features off the main ridges. Inspection of the actual peaks in the coefficients of the 2DFFT show bands at 1.4, 3.2, and 8.2 km wavelengths. Rotation of the topography 45° would cause a more symmetric appearance along the two diagonals because the ridge systems would affect each diagonal similarly.
2. Along the diagonal from lower left to upper right (southeast to northwest) the Fourier coefficients associated with longer wavelengths than 8 km are generally lower than the value of coefficients along the other diagonal. However, similar bands appear at about 2 km wavelength indicating the importance of the eight ridge features.
3. The band on each diagonal at about 8 km is due to the fact that southeast half the topography is more complex than the northwest half.
4. The higher frequency bands have about half the value for the Fourier coefficients than found at 8 km spacial wavelengths.
5. The coefficients are relatively symmetric about the diagonals demonstrating the predominate southeast-northwest and southwest-northeast orientation of the topography.

Obviously, comparisons with 2DFFT's of similar scale topography would be more useful in defining important parameters. It is interesting to note that the wavelength bands of 2 km and 8 km and measured nighttime drainage wind speeds of about 1.5 m/s found in a companion paper,² imply frequencies close to once per 27 minutes and 2 hours. These peaks were found in the autospectral analysis of wind

speeds in the region. This may be coincidental but should be investigated further in future analysis.

The next part of this section describes the results of digitizing photographic negatives of the acoustic sounder record near Thorn 7. Figure 3 shows the location of the acoustic sounder below the laser path used for optical anemometry. The difference in height of the beam and the acoustic sounder was about 30 meters. This is the height where the first usable signals for the acoustic sounder arrive. Figure 4 shows the acoustic sounder record for the period studied. This period was 12 hours beginning at midnight July 19, 1979. Each four hour segment was digitized with 200 line elements up to two hundred meters (i.e., 1 meter per line). The lines were chosen from computer 3-D data simulations like the one shown in Fig. 5. This helped to determine the edges of the picture and where the acoustic data began. Five line scans above 30 m were averaged at this point and the resulting data is plotted in Fig. 6 for the three four-hour segments. The height of the line is inversely proportional to the darkness of the photograph which in turn is proportional to the temperature structure parameter C_T . The periodic large single peaks correspond to timing marks on the record and were not included in the subsequent analysis. Ten minute average values of the inverse of the data plotted in Fig. 6 were calculated and compared to optical turbulence over the laser path shown in Fig. 7. The mathematical basis for this relationship relates the log amplitude variance of the laser intensity σ_I to C_T by the equations

$$\sigma_I^2 \propto k^{7/6} L^{11/6} C_n^2$$

and

$$C_n^2 \approx C_T^2$$

where L is the path length, k is the wavenumber and C_n is the index of refraction

structure parameter. These equations assume long path homogeneous isotropic atmosphere where humidity variations are negligible. Though differences appear in the comparison shown in Fig. 7, the correlation function shows a significant (F-test $< .01$) positive peak correlation (0.52) with higher values at night and late morning and simultaneous decrease in the early morning. These differences and the unsymmetric shape of the correlation function may be due in part to the fact that the optical turbulence is not always uniform along the path and the sounder is looking at only part of the path. For example, heating in the morning occurs first on one side of the path and then the other consistent with the hump on the left side of the correlation function.⁴ In the 1980 ASCOT experiment digitized data from the doppler acoustic sounder may improve this comparison.

CONCLUSIONS

Two dimensional Fourier transform of topographic relief have been shown to be useful to understanding the spacial variability in the Geyser's region of complex terrain. An interesting correspondence of regular spacial variability of topography with temporal regularities of wind speed variation was found which may or may not be coincidental. Digitized acoustic sounder records showed an expected relationship of optical turbulence to acoustic return signal strength.

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TOPOGRAPHY CONTOURS

G63

TOPOGRAPHY CONTOUR INTERVAL IS 50. METRES. MAXIMUM CONTOUR IS 1100.
 GRID ORIGIN UTM COORDINATES ARE: X= 520.0 KM, Y= 4285.1 KM, Z= 300 MASL.
 MESH INTERVALS ARE: DELX= 0.317 KM, DELY= 0.317 KM, DELZ= 50 METRES.

FIGURE 1. Digital topography height contours of 63.5 m resolution topography with only 317 m resolution horizontal and 50 m resolution vertical of 16 km square Geyser's region.

Two dimensional Fourier transform of Geyser's Terrain Contours
of Fourier coefficients

FIGURE 2. Contours of two-dimension fast Fourier transform of 63.5 m topography of region shown in Fig. 1.

SIDEVIEW OF LASER PATH IN GEYSER'S REGION

FIGURE 3. Cross section of laser path showing location of acoustic sounder about 30 m below near Thorn 7 part of the path.

FIGURE 4. Acoustic sounder record for 19 July 1979 for sounder shown in Fig. 3 starting at 0:00 PST and going to 12:00 PST.

FIGURE 5. Three dimensional plot of density for a section of photographic record of acoustic sounder data shown in Fig. 4.

FIGURE 6. Digital scans of three four-hour segments shown in Fig. 4 averaged 5 meters above 30 meter starting height for the sounder record.

TLAG CORR. ACOUSTIC VS SIG. OPT. H 7/19/79

FIGURE 7. Comparison of time histories of the photographic record function of C_2 from acoustic sounder record and σ_I from the optical turbulence of the laser beam with their correlation function plotted below.

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