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The Daytime Wind in Valleys Adjacent to the Great Salt Lake

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1

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1. Introduction

In 1986 Los Alamos National Laboratory was engaged by the US Army to study the meteorological aspects of emergency preparedness at several sites where toxic materials are stored and handled. The project included a series of tracer and meteorological field experiments in the vicinity of the Tooele Army Depot, 60 km southwest of Salt Lake City, Utah. These experiments generated a large data set for validating numerical simulations and for empirical analyses of the local meteorology. This paper discusses the main characteristics of the daytime, up-valley flow at the Utah site, including frequency of occurrence, horizontal and vertical structure, and temporal evolution. Some parameters controlling the variability in onset time for up-valley flow are identified, and an empirical forecasting scheme is discussed.

2. Site Description

Large topographic relief and the presence of two lakes are the most striking features of the region (Fig. 1). Valley and basin elevations range from 1284 m above sea level (ASL) in the Great Salt Lake basin to about 1520 m ASL in Rush Valley. The Great Salt Lake covers about 6200 km^2 but has a mean depth of only 7 m. Because it is shallow, surface water temperature stays in phase with daily mean air temperature throughout the year, and air-land differential heating occurs in all seasons.

We will discuss the daytime, near-surface flow in the Tooele and Rush valleys. These valleys lie between the Stansbury Mountains to the west and the Oquirrh Mountains to the east. Valley-to-peak relief generally exceeds 1500 m. The floor of Rush Valley is a semi-arid surface covered with a sparse cover of greasewoods, changing to sagebrush on the alluvial fans that run against the mountain slopes. Low, unsteady katabatic wind velocities at night show that the valley is not well drained. The valley's high elevation (relative to adjacent valleys), its semiarid surface, and its basin-like topography all strongly influence development of the daytime, up-valley wind.

Tooele Valley is lower in elevation and is open to the north. High velocities for katabatic flow at night show that it is well drained. The water table is close to the surface; mud flats and agricultural fields cover a large fraction of the northern portion of the valley.

Toxic materials are stored in the South Area of Rush Valley, tens of kilometers south of the population centers of Tooele Valley. Therefore, understanding the north-south exchange of air between these valley is of vital interest to those concerned with emergency preparedness. South Mountain forms a 450-m barrier to north-south flow; low points in this barrier are the narrow gap just north of Stockton and a higher pass on the mountain's western flank. The 700-mb pressure surface is normally just below the major summits of the region. We regard the wind at this level as representative of large-scale flow and will refer to it as the gradient wind. Sixty-six percent of the time during the summer months the 700-mb wind direction is in the range of 190° to 290° and has a mean scalar value of 7.5 m/s (Crutcher 1959).

3. Characteristics of the Daytime, Up-Valley Wind

Below 700 mb, a thermally driven, topographically controlled wind is often observed. This mesoscale circulation has a diurnal cycle forced by the differential atmospheric heating caused by sloping terrain and the contrasting thermal properties of land and water. The earliest detailed account of this circulation was given by Hawkes (1947), in which he discussed the diurnal wind systems in the Salt Lake and Dugway valleys and in the Ogden area. Smidy (1972) investigated the seasonal variation of near-surface winds but limited the study to winds in the Salt Lake Valley. More recently, the Wasatch Front Regional Council (1980), motivated by air pollution problems, conducted a study of the wind system in the Great Salt Lake basin. And Astling (1986), in a paper on mesoscale wind fields, discusses the influence of the Great Salt Lake on circulation in the Dugway Valley and along the southern shore of the lake. The Los Alamos study focused on the flow in Tooele and Rush valleys, valleys that have not received much attention in previous investigations. In this paper we discuss only the daytime phase of this circulation.

3.1 Wind Roses

The presence of a diurnal cycle in the near-surface flow is very evident in the frequency wind roses for direction shown in Figs. 2a-b. These plots are based on observations made 10 m above the ground at well-exposed sites. Statistics for stations 1 to 4 are based on a January-to-August sample in 1987; those for station 32 are based on a June and August sample in 1986.

In our sample, the surface flow is up-valley at least 50% of the time in the late afternoon (Fig. 2a). Investigators of the Wasatch Front Regional Council study (1980) concluded that local thermal forcing controlled the near-surface flow about 80% of the time, but most of their observations were made in and along the edge of the Great Salt Lake basin. The N to NE wind directions at station 4 and the very high frequency of NNW wind at station 32 suggest that the up-valley wind often penetrates 45 km inland from the southern shore of the Great Salt Lake. For reference, the nighttime wind direction pattern is shown in Fig. 2b; katabatic flows are evident throughout the region.

The frequent ENE wind through the low passes separating Cedar Valley from Rush Valley, opposite to the prevailing gradient wind, was an unexpected result and is discussed by Stone and Hoard (1990).

3.2 Structure of the Daytime, Up-Valley Wind

An example of the horizontal distribution of fully developed up-valley flow is shown in Fig. 3. The gradient wind during this period was 8 m/s out of the west. Typical near-surface velocities in Rush Valley are 4 to 7 m/s. During our experiments, up-valley flow was not observed in the southern portion of Rush Valley, so the maximum inland penetration of the flow is thought to be about 55 km. There, up-valley winds are part of a mesoscale circulation that has a horizontal scale of approximately 100 km. The lower portion of this circulation begins far offshore, a fact well understood by local wind surfers and sailboat enthusiasts.

Figure 4 shows vertical profiles of wind, temperature, and moisture in up-valley flow conditions in central

Rush Valley. In this example, a 300-m-deep up-valley wind arrived just after sunset. The temperature profile is particularly interesting because it shows the effects of diabatic cooling near the surface. The 1700 MST sounding in Salt Lake Valley for this day showed onshore flow to a depth of 500 m, and gradient winds were SSW at 3 m/s. Another set of late afternoon profiles shows a 500-m-deep up-valley flow in Rush Valley with onshore flow to a depth of 700 m above Salt Lake City. Although the soundings above Rush Valley and Salt Lake City are not concurrent, evidence of a sloping frontal interface (1:200 here) has been noted by others (e.g., Frizzola and Fisher 1963).

3.3 Evolution of the Up-Valley Wind

The onset of up-valley flow at any given station is usually clearly indicated by an abrupt change in wind direction and an increase in wind speed. After onset, the temperature trace often shows a slowdown or cessation of daytime heating. Generally, however, the temperature record by itself is not a good indicator of onset time.

When onset times are plotted as a function of distance south of the lake, we find that up-valley flow progresses southward, with the speed of advance increasing with distance from the lake (Fig. 5). The figure is based on a small sample from July and August 1987, but the sample is large enough to show that onset time at a given station is quite variable. We also note that the rate of advance of the wind reversal is variable.

Some of the variability in onset time appears to be caused by changes in the soil moisture content in Rush Valley. The solid lines in Fig. 5 show onset times for 2 days when the soil moisture was 7% in the upper 10-cm layer, and the dashed lines show onset times for 2 days when the soil moisture was 15%. The gradient wind was less than 8 m/s and was mostly westerly on all four days, so we tentatively conclude that changes in the horizontal distribution of soil moisture can strongly influence arrival time of the up-valley wind through its role in the surface energy balance. The sensible heat flux density over the Rush Valley surface during the "dry" period was observed to be approximately twice that during the "wet" period. Modeling studies by Ookouchi et al. (1984) show that mesoscale circulations are indeed very sensitive to the horizontal distribution of available soil moisture.

The advance of the up-valley wind behaves as a front, at least in the sense that a line of convergence moves southward. In Tooele Valley the rate of advance is typically 1 m/s, much slower than wind speeds behind the front, which are typically 3 to 5 m/s. Because the front moves slowly and the transition (at a station) is rapid, the frontal zone must be narrow; in Tooele Valley we estimate it to be 1 to 2 km wide. By the time the front reaches Rush Valley, its speed increases to 6 m/s, which is comparable to normal wind speeds for the up-valley wind in that area.

As mentioned above, a temperature drop is usually not observed with passage of the front. However, in rare cases, such as late arrivals under the influence of strong southerly gradient winds, a drop of 2 to 4°C is observed. Normally, diabatic heating of the cool lake air over the land is so rapid that it more than compensates for temperature advection.

Probability density histograms are used to characterize the variability in onset time of the up-valley wind at stations 1 and 2 (Fig. 6). The sample for each station includes all cases from the January-to-August dataset for which instrumentation was working reliably and the wind direction reversal had the signature of local thermal forcing. Histograms for cessation times are also shown; cessation of the up-valley wind marks the onset of katabatic down-valley flow. These histograms illustrate some interesting features:

(1) At station 2, the distributions are broader and median onset and cessation times are later than those at station 1. The distributions for onset time for up-valley flow cover almost the whole range of the daytime heating cycle, therefore they would not be very helpful for forecasting wind direction reversal times.

(2) Distributions for cessation time at both stations are much narrower. A tentative explanation for this lies in the very different states of the valley atmospheres at sunrise and sunset. By late afternoon the atmosphere is in a nearly adiabatic state. Any local cooling of air near sloping terrain quickly results in the baroclinity that drives katabatic flow. Also, vertical mixing subsides in late afternoon, which decouples the surface flow from the variable influence of the gradient wind.

At sunrise, on the other hand, katabatic flows have concentrated the diabatic cooling that has occurred during the night. In poorly drained valleys and basins like

Rush Valley, deep ground-based temperature inversions form, and these have to be destroyed before a low-level, N-S temperature gradient can develop. Also, during the morning hours soil moisture availability strongly influences the sensible heat flux, and the influence of the gradient wind strengthens as vertical mixing begins.

(3) It is not unusual for up-valley flow in Rush Valley to continue for an hour or two after sunset, even though down-valley flow is under way in Tooele Valley.

At station 1, we observed a complete cycle of the thermally forced wind on 70% of the days for which reliable data were available; the sample included data from April to early August 1987. At station 2, we observed a 50% occurrence of the cycle; the sample included data from late February to early August 1987.

Many of the features of the up-valley wind south of the Great Salt Lake are similar to those found in the literature on lake and sea breezes. Wind speed, depth, frontal advance speeds, inland penetration, and onset time variability are all in the range found in the literature (see, for example, Simpson et al. 1977; Lyons 1972; Keen and Lyon 1978; Moroz 1967; and other references cited in Atkinson 1981). At the Utah site one does not see the effect of the Coriolis force, which is often observed at other sites; topographical constraints on the flow may account for this.

4. Forecasting the Onset of the Up-Valley Wind

Many sites with the potential for accidental releases of toxic materials are located in areas where diurnal wind direction reversals occur because of local heating and cooling. Accurate methods for forecasting the time of these wind reversals could provide critical information for responding to an emergency.

In principle, numerical simulations of the flow should provide the needed information. Models, both analytical and numerical, have been used extensively to study thermally driven mesoscale flows, and the results are generally impressive (see references cited by Atkinson 1981). However, most of these investigations involve case studies with comparisons with limited datasets, or they involve parametric studies with no reference to observations. To our knowledge, Rye's (1989) study is the only one that addresses the problem of forecasting the onset time of a sea breeze in a routine application (providing forecasts during the 1986-87 America's Cup yacht races). Although the model showed skill in other respects, Rye concluded that it showed almost no skill in forecasting onset time. Others, for example, Estoque et al. (1976) and Moroz (1967), have found that simulations underestimate intensity and inland penetration of the lake breezes. Underestimations also were found in preliminary model runs of the daytime flow at the Utah site (results unpublished).

The requirements for accurate routine simulations of the flow in Rush and Tooele valleys are formidable. Site complexity requires a three-dimensional calculation over a large domain. Sharp temperature and velocity gradients, as well as small but important topographic features, require a small grid spacing, perhaps 4 km or less. Initial and boundary conditions must be specified in detail. Specifically, parameters that affect the surface energy balance, such as the spatial distributions of soil moisture, snow, and clouds, would have to be specified. Time variation of the large-scale flow would also have to be incorporated. The expense and complexity of providing this kind of information are perhaps as large a challenge as doing the calculation. At the present time, in our opinion, numerical simulations, using standard initialization schemes, cannot predict onset times with sufficient accuracy for emergency response applications.

Results of an analysis of 7 months of data from stations 1 and 2 suggest that an empirical approach to the forecast problem may be more promising. Here we adopt the view that most of the variance in the arrival time can be explained by a few important, easily observed parameters.

Biggs and Graves (1962) adopted this approach in a study of the lake breeze off of Lake Erie. They used dimensional arguments to identify important dimensionless numbers that govern the lake breeze phenomenon. The ratio of the two most important numbers defined a lake breeze index, $\sigma = V^2/\Delta T$. Here, for simplicity, we have included the constants in the index. V is wind speed, irrespective of direction, ΔT is the difference between the maximum air temperature T_a and lake water surface temperature. Both V and T_a are supposed to be representative of conditions near the surface outside the influence of the lake. The index is interpreted as a ratio of inertial to buoyancy forces; when used in a diagnostic sense, the authors found that a critical value of σ separates lake-breeze days from non-lake-breeze days about 90% of the time. In other words, for a given V, a lake breeze occurred when ΔT exceeded a critical value of V^2/σ. Walsh (1974) considered the index concept within the context of his analytical model of the sea breeze and found theoretical support for the linear relationship between V^2 and a critical ΔT.

Lyons (1972) applied the index concept to a study of the lake breeze across the southwest coast of Lake Michigan. He was also able to find a critical value for σ that discriminated between lake-breeze and non-lake-breeze days about 90% of the time. However, attempts to correlate σ with onset time of the lake breeze and inland penetration were disappointing. The index apparently has no prognostic value. This is not surprising, since the element of time is not involved.

If the gradient wind is weak, the onset of the up-valley wind should depend mainly on the rate of heating of the valley atmospheres and the amount of heat the system needs to develop the critical N-S temperature gradient. (In our notation, a positive N-S gradient means that temperature increases toward the south.) We have already noted (in Fig. 5) that changes in the soil moisture in Rush Valley modulate the sensible heat flux and strongly influence onset time. We now consider the amount of energy the system requires to develop the baroclinity needed for the up-valley flow. The following observations give some clues about this energy requirement.

(1) The diurnal temperature range in Rush Valley is larger than in Tooele Valley. At station 2, the minimum temperature averages 5.2°C colder, and the maximum temperature averages 3.4°C warmer than at station 1.

In the daytime, a larger sensible heat flux over the Rush Valley's surface and the valley's smaller volume-to-area ratio contribute to the excess heating and development of a positive N-S temperature gradient. During the night, cold air accumulates in the bottom of Rush Valley because it is not well drained. Tooele Valley, on the other hand, is well drained to the north, and the nocturnal temperature inversions are less intense. Therefore the presunrise N-S temperature gradient is always negative.

(2) During dry, light-wind conditions, the presunrise temperature difference is larger than average. In such cases the N-S temperature gradient changes sign later in the day, and the up-valley wind arrives late.

(3) When cloudy and/or windy conditions prevail during the night, the presunrise N-S temperature difference is small. In such cases, the temperature gradient usually changes sign early in the morning, and the up-valley wind arrives early.

(4) The onset of up-valley flow at station 2 occurs 1 to 3 hours after the N-S temperature gradient becomes positive. Although this time delay is variable, the time of the gradient's sign change appears to have some forecast value.

(5) Typically, the temperature in Rush Valley climbs very rapidly in the early morning hours, exceeds that in Tooele Valley about 4 hours after sunrise. The temperature gradient usually reaches its maximum value around noon, stays relatively constant for several hours, and then slowly diminishes during the late afternoon. The average daytime N-S temperature gradient of 3.4°C/10 km is an order of magnitude greater than typical gradients across synoptic scale fronts. On days when a positive N-S temperature gradient did not develop, up-valley winds did not occur.

(6) There is no correlation between the magnitude of the N-S temperature gradient and arrival time of the up-valley wind, frontal speed, or its maximum velocity.

Observations 1 to 3 suggest that the energy deficit of Rush Valley, relative to that in Tooele Valley, plays a role in determining the onset of the up-valley wind. For a given sensible heat flux, the larger this relative deficit, the longer it takes to develop the baroclinity required to drive up-valley flow.

Figure 7 is a scatter plot of onset time to of up-valley flow at station 2 vs the presunrise temperature difference ΔT between stations 1 and 2. Unfortunately the dataset for which both stations were operating properly contains only 28 cases spanning April to early August. To the best of our knowledge, the solid dots, about 70% of the cases, represent the "normal" range of conditions. The open circles are cases in which the Rush Valley surface was unusually wet or when strong gradient winds were known to exist. Had we assumed a linear relation between to and ΔT, shown by the straight line, and used ΔT as a predictor, we could have forecast onset time with a standard error of 57 min under "normal" conditions; using the median onset time as a predictor, the standard error would have been 98 min.

Theoretically we expect the relationship between t_0 and ΔT to be nonlinear. We have been able to show (in work in progress) that the time t it takes to develop a positive, critical N-S temperature difference ΔT_{crt} is given by the following expression:

$$t = \left[\frac{\overline{\rho c_p}(\Delta T_{crt} + \Delta T)}{\left(\frac{A}{V}\right)_R \alpha_R - \left(\frac{A}{V}\right)_T \alpha_T} \right]^{1/2}$$

where ΔT is the presunrise difference $(T_T - T_R)$; $\bar{c}_p$ and $\bar{\rho}$ are average values of air specific heat and density; $\frac{A}{V}$ is a ratio of valley top area to the volume of air affected by the heating; α is the time rate of change of the sensible heat flux density; and the subscripts R and T refer to Rush and Tooele valleys, respectively. In this preliminary work, we modeled the presunrise potential temperature inversions as layers with constant lapse rates, different in the two valleys. And we assumed a constant slope to the sensible heat flux curve, an assumption usually justified during the first part of the heating cycle. We are currently investigating the utility of incorporating this type of modeling into an empirically based forecasting scheme. Such a scheme would account for both the energy required to develop ΔT_{crt} and the rate of energy deposition.

Our few observations regarding gradient wind effects are in accord with the consensus one finds in the lake/sea breeze literature. Namely, onshore gradient winds (up-valley at our site) result in earlier arrivals, and offshore winds (down-valley at our site) delay the arrival. However, the number of cases with significant along-valley gradient wind components is so few, we were not able to study gradient wind effects, a deficiency our study shares with many other observational studies of the lake and sea breezes.

5. Conclusions

We have used 7 months of data-weighted toward the summer months-to characterize the daytime phase of the diurnal mesoscale wind circulation in the valley south of the Great Salt Lake. At this topographically complex site, surface wind direction reversals occur about 70% of the time as katabatic down-valley flow gives way to anabatic up-valley flow. Once established, the up-valley flow penetrates about 50 km inland, where it has a depth of 300 to 500 m and a near-surface velocity of 4 to 7 m/s. Driven by a N-S temperature gradient an order of magnitude larger than synoptic values, the wind advances southward during the day with frontal speed increasing from 1 m/s near the lake to 6 m/s near the end of its range. Onset time at a particular station is extremely variable; cessation time is less so.

In the context of emergency prepared-ness, accurate predictions of the onset time of up-valley flow could be vital. It is observed to be strongly influenced by the sensible heat flux, presunrise temperature inversion strength in the upper valley, and the gradient wind. Onset of the up-valley wind at inland sites is delayed when strong inland temperature inversions exist, the soil is wet (which reduces the sensible heating), the gradient wind is offshore, or by a combination of these factors. It may be possible to relate the onset time to an index based on these parameters. A pre-liminary analysis, accounting for only the temperature-inversion effects, suggests that it may possible to predict the onset time with a standard error smaller than 60 min at a site 25 km inland.

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Figure captions

Fig. 1. Map showing the main topographical features and population centers of the study area.

Fig. 2. Frequency wind roses for direction for (a) late afternoon and (b) early morning. The length of a spoke represents the percentage of the time, within the interval shown, that the wind blows from the indicated direction. Ring interval is 5%.

Fig. 3. Fully developed up-valley flow. Vector mean surface winds, 10 m above the ground, averaged from 14 MST to 18 MST, August 4, 1987. Terrain contour interval is 275 m. Velocity scale in upper left corner.

Fig. 4. Vertical profiles of wind direction, wind speed, water vapor mixing ratio, and temperature observed with a tethersonde in central Rush Valley, 2005-2046 MST, July 23, 1987.

Fig. 5. Onset time of the up-valley wind vs distance south of the Great Salt Lake. Dashed lines show 2 days when the Rush Valley surface was wet; solid lines show 2 days when the surface was dry.

Fig. 6. Probability density histograms for onset time (t_o) and cessation time (tc) of the up-valley wind at stations 1 and 2.

Fig. 7. Scatter plot of the onset time, t_o, for the up-valley wind at station 2 vs the temperature difference, ΔT, between Tooele and Rush valleys just before sunrise. The straight line, drawn by eye, through the solid points is thought to represent "normal" conditions. Open circles show effects of strong gradient winds and/or high soil moisture in Rush Valley.

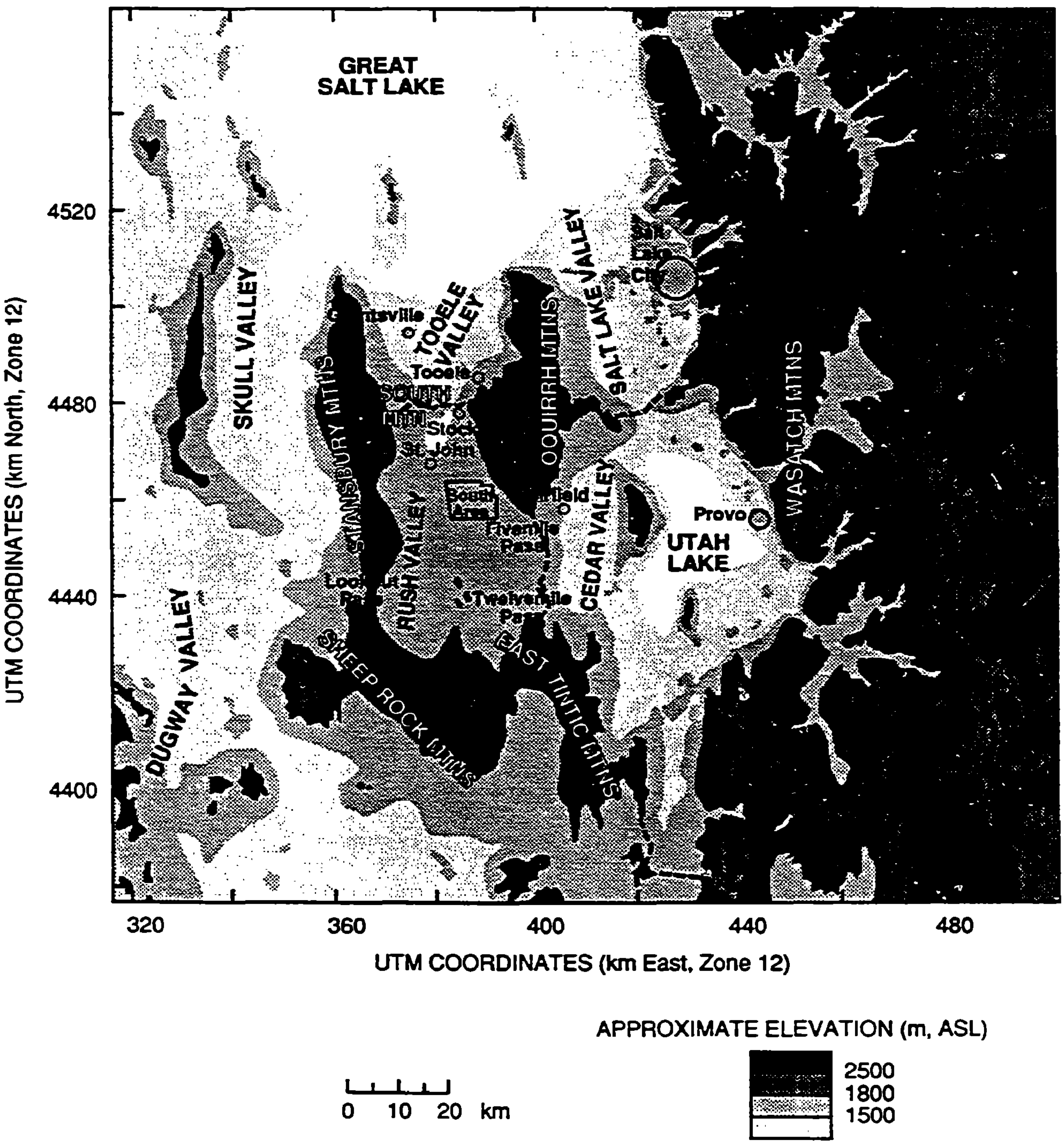
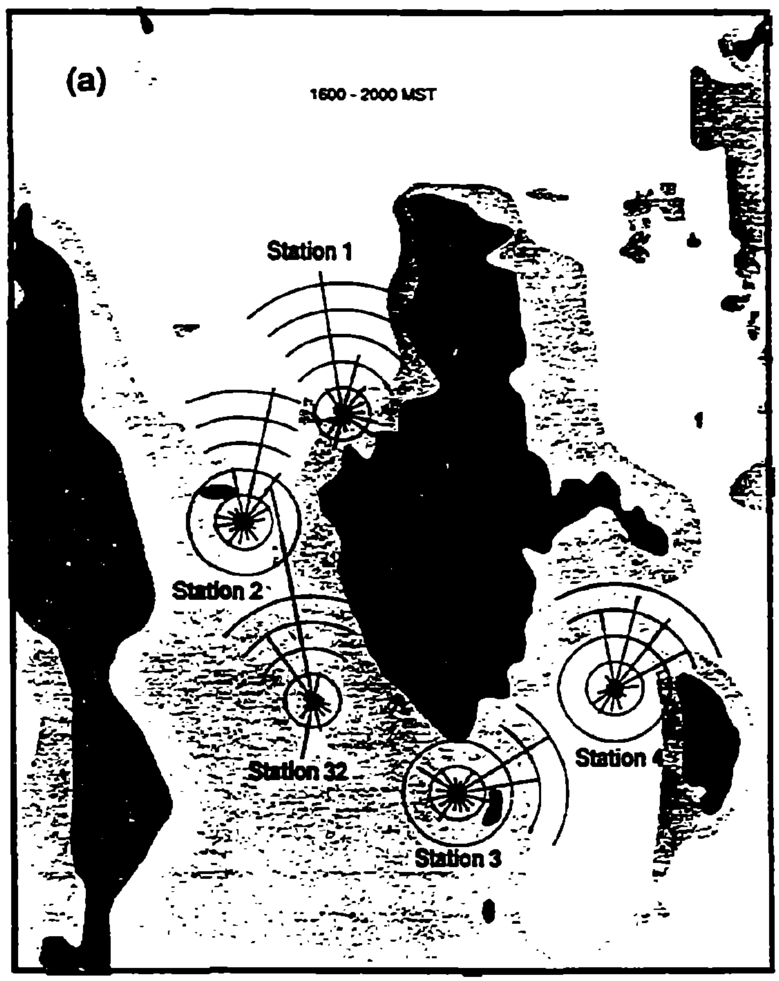
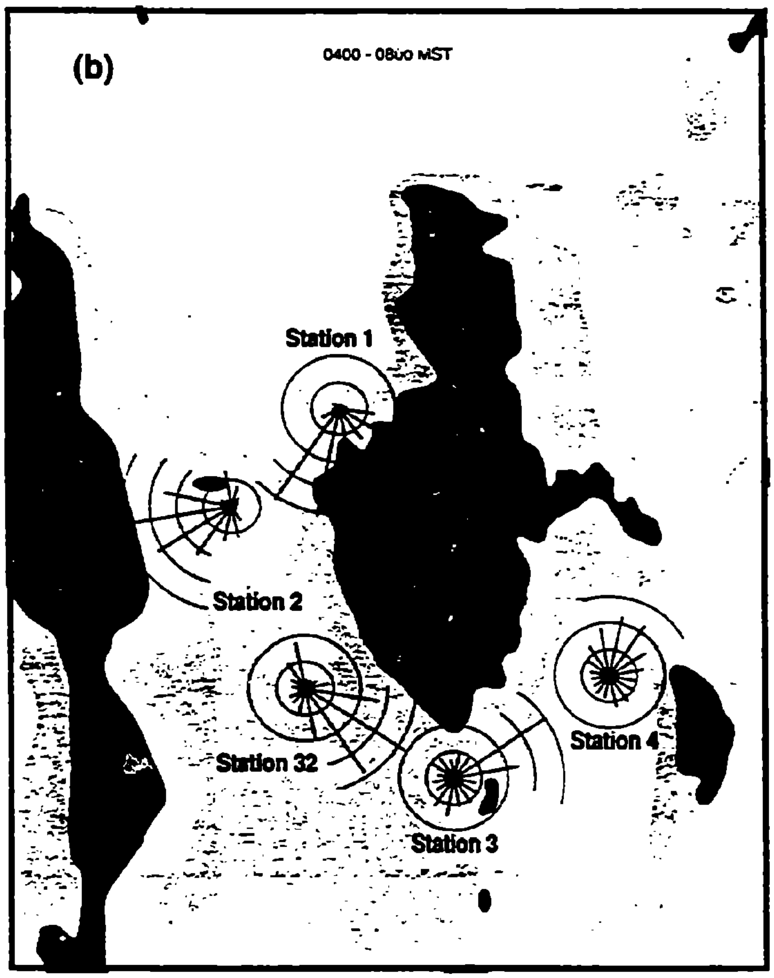

GREAT SALT LAKE
SKULL VALLEY
DUGWAY VALLEY
STANSBURY MTNS
TOOELE VALLEY
Tooele
Stockton
St. John
RUSH VALLEY
Fivemile Pass
Lookout Pass
Twelvemile Pass
SHEEP ROCK MTNS
EAST TINTIC MTNS
OQUIRRH MTNS
SALT LAKE VALLEY
Salt Lake City
CEDAR VALLEY
Fairfield
UTAH LAKE
Provo
WASATCH MTNS
4520
4480
4440
4400
320
360
400
440
480
UTM COORDINATES (km North, Zone 12)
UTM COORDINATES (km East, Zone 12)
APPROXIMATE ELEVATION (m, ASL)
2500
1800
1500
0 10 20 km

Fig. 1.



(a)
1600 - 2000 MST
Station 1
Station 2
Station 32
Station 3
Station 4

Fig. 2a



(b)
0400 - 0800 MST
Station 1
Station 2
Station 32
Station 3
Station 4

Fig 2b

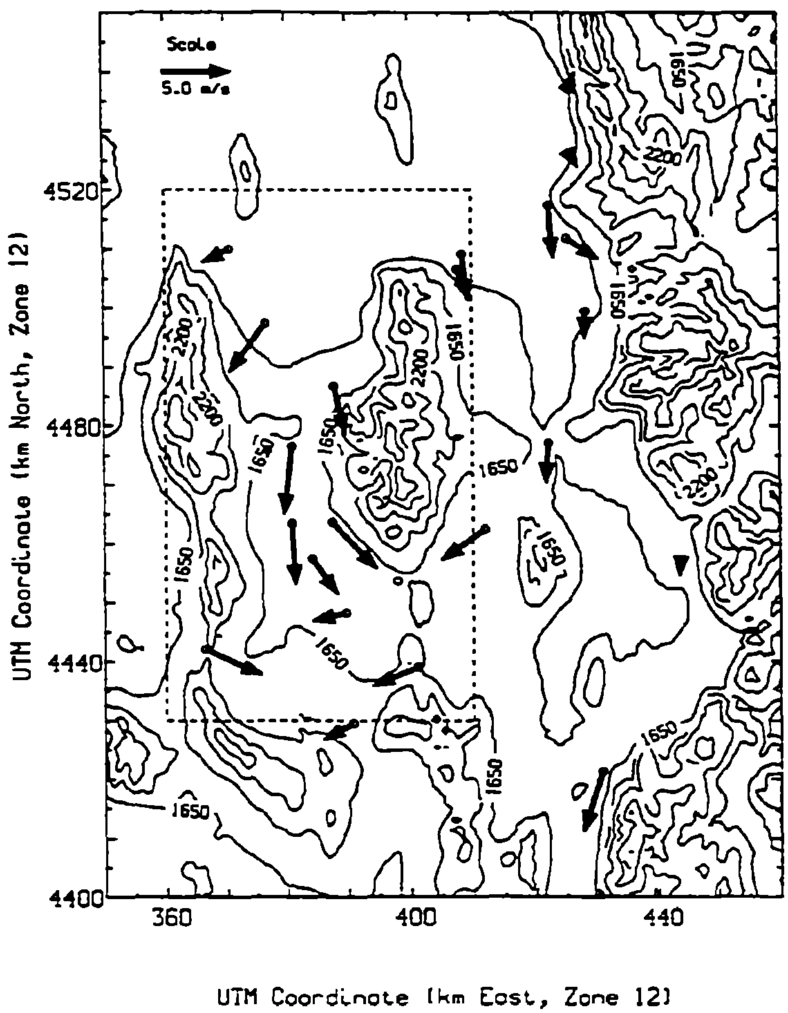

Scale
5.0 m/s
4520
4480
4440
4400
360
400
440
UTM Coordinate (km North, Zone 12)
UTM Coordinate (km East, Zone 12)
1650
2200

Fig. 3

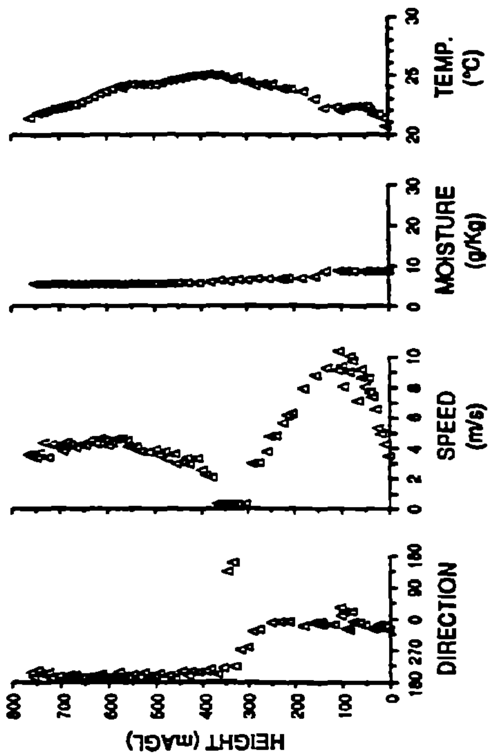


Fig. 4.

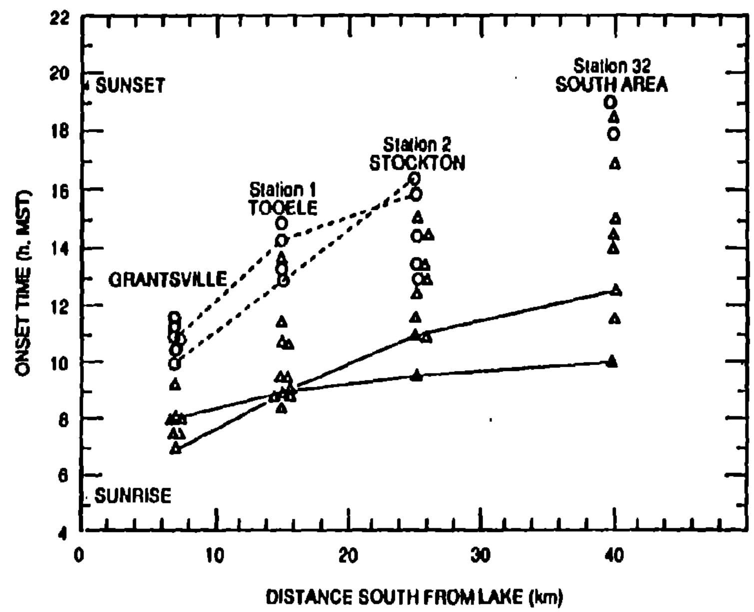
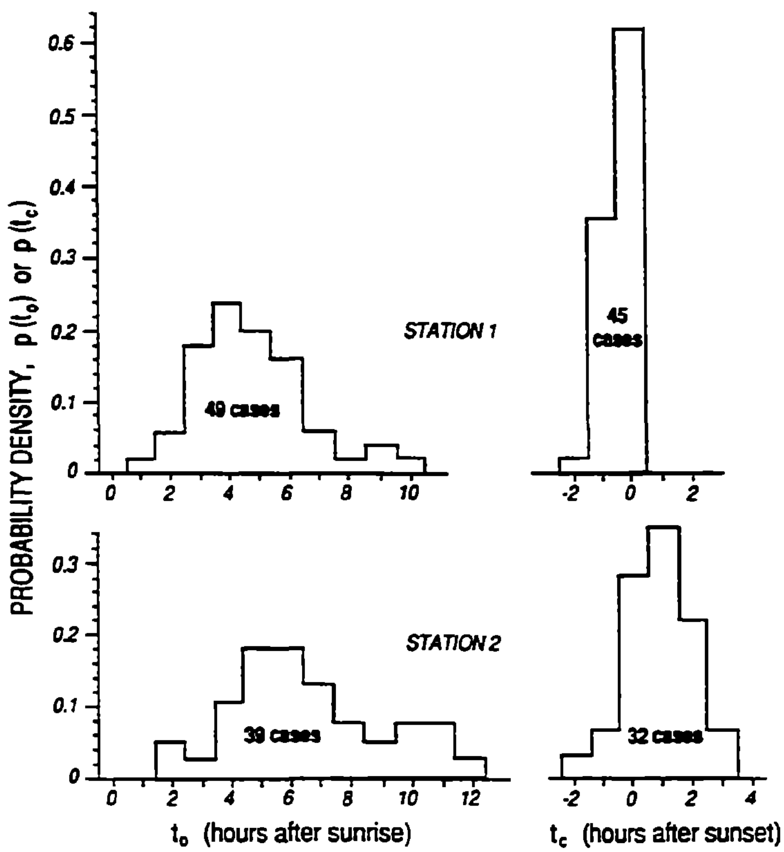

Station 32
SOUTH AREA
Station 2
STOCKTON
Station 1
TOOELE
GRANTSVILLE
SUNSET
SUNRISE
ONSET TIME (h. MST)
DISTANCE SOUTH FROM LAKE (km)
22
20
18
16
14
12
10
8
6
4
0
10
20
30
40

Fig. 5



PROBABILITY DENSITY, p (t_o) or p (t_c)
0.6
0.5
0.4
0.3
0.2
0.1
0
STATION 1
49 cases
45 cases
0 2 4 6 8 10
-2 0 2
0.3
0.2
0.1
0
STATION 2
39 cases
32 cases
0 2 4 6 8 10 12
-2 0 2 4
t_o (hours after sunrise)
t_c (hours after sunset)

Fig 6.

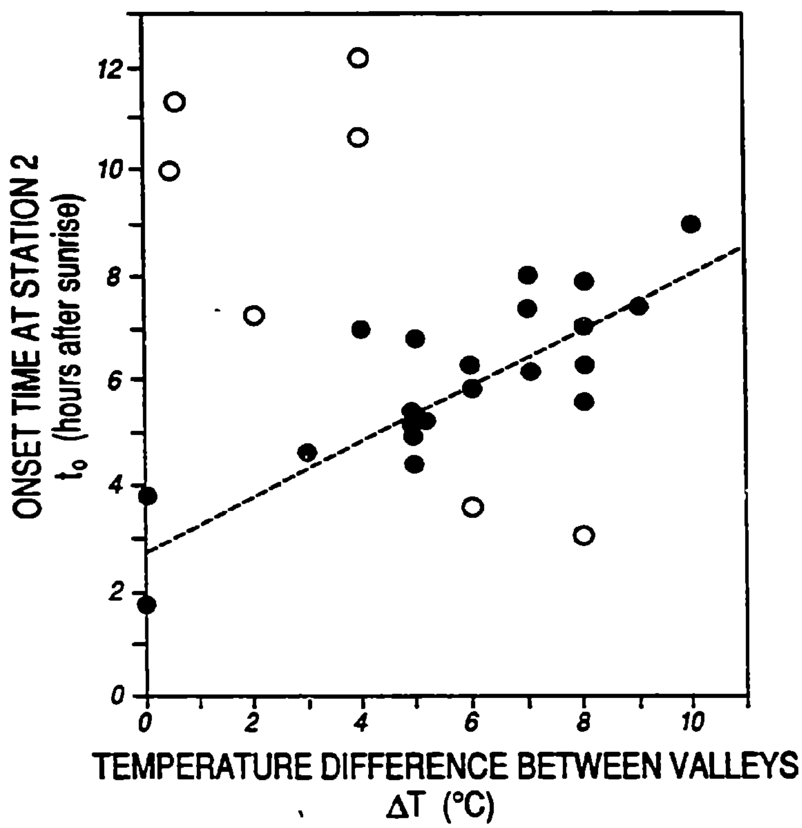

ONSET TIME AT STATION 2
t_o (hours after sunrise)
12
10
8
6
4
2
0
0
2
4
6
8
10
TEMPERATURE DIFFERENCE BETWEEN VALLEYS
ΔT (°C)

Fig. 7.