A Proposed Index for Mesoscale Activity

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ABSTRACT

The diurnal and interdiurnal wind variabilities are defined in terms of the "relative gustiness"—$\sigma_v/\bar{V}$. The proportion α between the diurnal and the interdiurnal variabilities thus defined is suggested as a useful index in mesoscale studies. The α index is calculated in more than 30 stations in the Israel area during summer and winter, and is shown as a simple tool for the identification of climatic-geographic regions in which:

i) The variability due to the changes of the large-scale pressure gradients may or may not be ignored in mesoscale models. An α index larger than 1 indicates the dominance of the diurnal processes over the interdiurnal processes in producing wind variability, in which case large-scale effects are relatively small.

ii) Hodograph representation for the average diurnal wind variation is or is not a faithful picture of the actual wind observations on a particular day. An α index larger than 1 suggests a more faithful representation of the wind hodograph.

As expected, the α index is larger than 1 during summer in nearly all stations in Israel while during winter it is less than 1.

The values of α during summer are very high. They are generally between 1.5 and 2.5 but in the Jordan Valley, Arava and the Negev desert α indices reach 3 or even 4–5, indicating both vigorous mesoscale forcing due to channeling and/or differential heating and the reduced effect of the migrating cyclones. In the winter only few southern stations get α-values slightly larger than 1, indicating the general dominance of the large-scale (synoptic) variability relative to the locally induced mesoscale variability.

1. Introduction

In recent years much interest has been given to mesoscale modeling and, in particular, to circulations induced by boundary layer forcing. Pielke (1981, p. 290) refers to the circulations that are primarily forced by surface inhomogeneities as terrain-induced mesoscale systems. It has been established by numerous successful simulations that these models show a very high potential for short-range forecasting and local weather prediction. To give only a few examples of these simulations, we will mention those of Pielke (1974), Colton (1976), Anthes and Warner (1978) and Alpert and Neumann (1983). It has also been shown that mesoscale models predict well diurnal wind variabilities over complex terrain, and some investigators even succeeded in wind predictions with only one-layer simulations in which surface inhomogeneities play the most important role. For examples of the latter, reference is made to Lavoie (1974), Danard (1977) and Mass and Dempsey (1985).

An important question regarding the prediction of the diurnal wind changes is, "What are the basic processes which contribute to the wind variability over a diurnal cycle?" Of special interest is the relative contribution of the differential heating to the diurnal wind variability, which will be referred to later as the "mesoscale contribution." In this study we first estimate the variability due to the diurnal stratification changes over homogenous terrain, which we may crudely refer to as the microscale variability. Second, the interdiurnal wind variability, referred to as the large-scale contribution, is calculated. All wind variabilities are defined in terms of an extended gustiness; i.e., a normalized wind variability.

In the next section we define an α index for mesoscale activity as the proportion of the diurnal to the interdiurnal wind variabilities in terms of the "extended gustiness." This is done after the microscale contribution is estimated to be relatively small, i.e., about 0.11–0.16. The values and the distributions of the diurnal and interdiurnal wind variabilities in over 30 stations in the Israel region during July and January 1978 are discussed in Section 3. In Section 4 the α index is calculated and its importance for mesoscale modeling is discussed. Section 5 illustrates the importance of the α index in estimating the average hodograph representativeness of the actual wind measurements. The sixth and last section summarizes and discusses the important conclusions.

2. Index for mesoscale activity

It is customary to define the wind gustiness or the turbulence intensity as the ratio of the root-meansquares of the eddy velocities to the mean wind speed; see, e.g., *Glossary of Meteorology* (1959). For the east–west wind component u the turbulence intensity is given by

$$I_x = (\overline{u'^2})^{1/2}/\bar{V} = \sigma_u/\bar{V}, \quad (1)$$

where $\bar{V}$ is the mean wind speed. The turbulence intensity is sometimes referred to also as the "relative gustiness." In general, x is chosen as the direction of the average wind vector and I_x is referred to as the longitudinal or downwind component. Sutton (1953) finds that nearly all values of I_x for a grass field 1–2 cm high lie between 0.1 and 0.2. Also, I_x does not vary appreciably with height (near the surface) and is independent of the mean velocity to a moderate degree of approximation. Swanson and Cramer (1965) have calculated the turbulence intensities based on the White Sands Tower data, and have shown that they decrease with height in all thermal stratifications, tend to be inversely proportional to the mean wind speed and decrease with increasing stability. However, most of the turbulence intensities at the height of ~12 meters are in the range of 0.15–0.25. The calculations were done for 10-minute periods, i.e., $\bar{V}$ is the wind average over 10 minutes. Following Lumley and Panofsky (1964) the relative gustiness in neutral air could be approximated by

$$I_x = Ck/\ln(z/z_0), \quad (2)$$

where C is a constant independent of height ($C \approx 2.5$), k is von Kármán's constant and z_0 the roughness length. Taking $C \approx 2.5$, one finds that I_x, at a certain height z above the surface, is solely determined by the roughness length and is approximated by $[\ln(z/z_0)]^{-1}$. However, the value of C varies and Lumley and Panofsky (1964), who give some examples of the value of C (2.1–2.9), suggest that C seems to vary with *terrain*. This feature of the variance of the longitudinal velocity, which is found even on the short turbulent time scales (typically ~10 min averaging time interval), will be shown later to be relevant when averaging over longer time intervals. Also, the variance of the longitudinal velocity is not too sensitive to changes in stability as are, for example, the lateral velocities. This was illustrated by Lumley and Panofsky (1964, p. 154) and is also of importance to the present study.

The typical averaging time interval in turbulent (microscale) studies is of the order of a few minutes. However, we would like to consider longer time intervals and define analogously the normalized diurnal wind variability I_B as:

$$I_B = \frac{\sigma_B}{u_j} = [\sum_{i=1}^{24} (u_{ij} - u_j)^2]^{1/2}/(24^{1/2} \cdot u_j), \quad (3)$$

where u_{ij} is the longitudinal component of the wind (i.e., wind intensity) at hour i and day j, and

$$u_j = (\sum_{i=1}^{24} u_{ij})/24$$

is the average diurnal wind in the longitudinal direction for the jth day. The average value of I_B over a period of N days is given by

$$\bar{I}_B = \frac{1}{N} \sum_{j=1}^{N} I_B. \quad (4)$$

Here I_B is also referred to as the "diurnal relative gustiness." Similarly, we can define the normalized interdiurnal wind variability I_A as

$$I_A = \frac{\sigma_A}{u} = [\sum_{j=1}^{N} (u_j - \bar{u})^2]^{1/2}/(N^{1/2} \cdot \bar{u}), \quad (5)$$

where

$$\bar{u} = (\sum_{j=1}^{N} u_j)/N.$$

In the next section we shall present the computed values of I_A and $\bar{I}_B$ and show that

$$\alpha = \bar{I}_B/I_A \quad (6)$$

might be a useful index in mesoscale studies. Also, α will be suggested as an index for the identification of climatic–geographic regions in which the average diurnal wind hodograph represents faithfully the actual diurnal wind changes.

Furthermore, in most mesoscale models, e.g., Mahrer and Pielke (1978), Anthes and Warner (1978) and Alpert *et al.* (1982), the large-scale pressure gradient is assumed to be constant through at least one or two diurnal simulations. The extent to which such an assumption is appropriate could be determined by the value of the pertinent index α; i.e., for $\alpha > 1$ the diurnal wind variability is dominant over the synoptic-scale interdiurnal variability and vice versa.

When the synoptic-scale interdiurnal variability becomes relatively large ($\alpha < 1$), neglect of the transient effect of the large-scale pressure gradient, which actually occurs in most mesoscale modeling, could hardly be justified. For a discussion of the importance of the transient large-scale effect on the sea breeze see, e.g., Haurwitz (1950).

Of course, the diurnal variability defined by I_B—Eq. (3)—cannot be identified solely with mesoscale variability because there is evidently an important "microscale" contribution to the diurnal variability due to the diurnal stratification changes.[1] In order to get a first-order estimation to the latter contribution we adopt the Wipperman *et al.* (1973) results for the

[1] The name "microscale" does not mean that mesoscale models are not capable of simulating such a contribution. The term "microscale" is only used for the separation of wind variability that is independent of any horizontal scale from that caused by horizontal inhomogeneities and therefore associated with some horizontal scale.

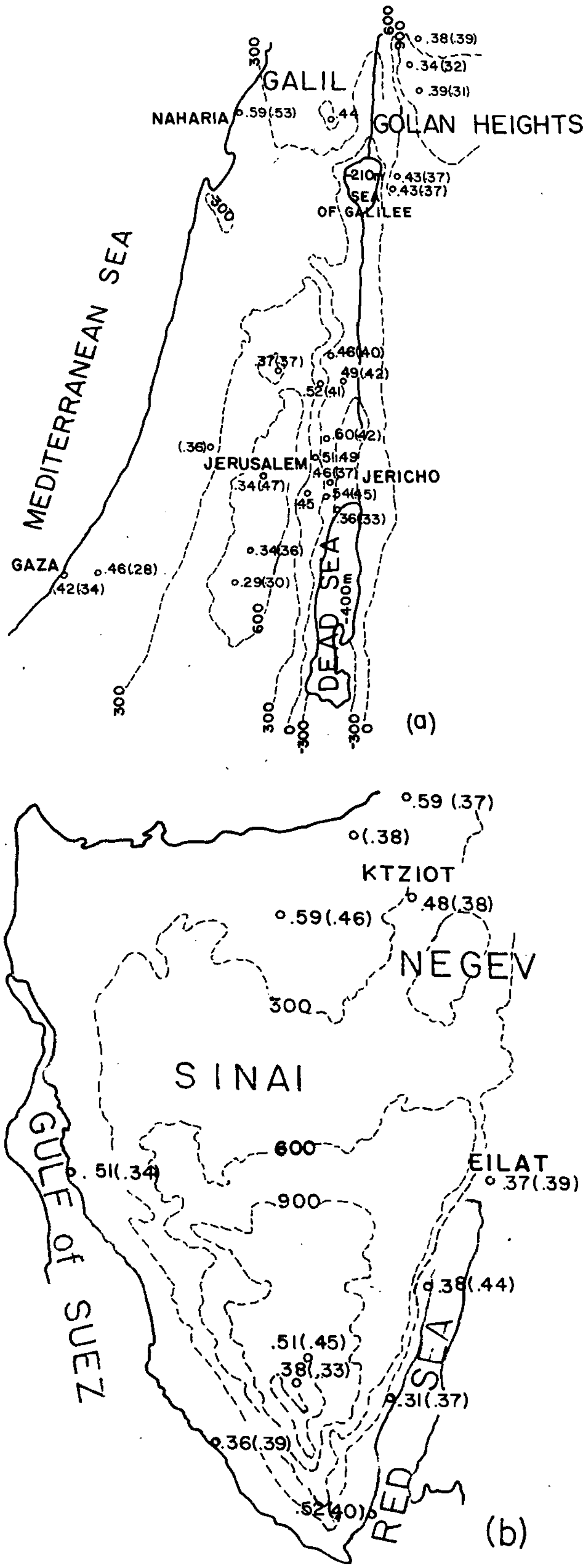

FIG. 1. The monthly-averaged normalized diurnal variabilities $\bar{I}_B$ [see Eq. (4)] for some stations in the Israel region for July and January 1978. Dashed lines represent topographic contours with an interval of 300 m. The values in parentheses represent the January numbers.

diurnal wind variation of the nonstationary, horizontally homogeneous, planetary boundary layer model. The model solves the following momentum Eqs:

$$\frac{\partial u}{\partial t} - f(v - v_g) = \frac{\partial}{\partial z}\left(\frac{\tau_x}{\rho}\right); \quad \frac{\tau_x}{\rho} = k_m \frac{\partial u}{\partial z} \tag{7a}$$

$$\frac{\partial v}{\partial t} - f(u + u_g) = \frac{\partial}{\partial z}\left(\frac{\tau_y}{\rho}\right); \quad \frac{\tau_y}{\rho} = k_m \frac{\partial v}{\partial z}, \tag{7b}$$

where: $u(u_g)$ and $v(v_g)$ are the averaged horizontal (geostrophic) velocity components, τ_x and τ_y the components of Reynolds stress, f is the Coriolis parameter, ρ the density, t the time and z the vertical coordinate. The momentum eddy viscosity k_m is a function of the thermal stratification. The latter is deduced diurnally from a realistically prescribed surface heat flux. Wipperman *et al.* (1973, p. 46) list the values of parameters which have been used for the simulation of the diurnal variation of the boundary layer. Following their results (see their Fig. 7) for the nondimensionalized magnitude of the horizontal wind, $|\mathbf{V}|/|V_g|$ as function of time and of the nondimensionalized height $z_* = z/(ku_*/f)$, we have calculated the value of I_B at $z_* = 0.01$, corresponding to the approximate height of 12 meters, to be 0.11. This value is rather close to that predicted for the short turbulent scale in neutral air (~0.14 for the same roughness length of 10^{-2} m); see Eq. (2). In conclusion, the value of 0.11 for I_B will be used as an estimate of the microscale contribution to I_B. A value quite close to that, 0.16, will be reached by analysis of the O'Neill observations in the next section.

3. Diurnal and interdiurnal wind variabilities in Israel region—Calculation of I_B and I_A

The hourly wind data was supplied by the IMS (Israel Meteorological Service) and was based upon one-hour averages from continuous wind recorders. The average wind was calculated through the hourly course of the wind. All of the surface wind observations were reduced to the ten meter height by using Hellman's formula; see, e.g., Sellers (1967, p. 152). Figure 1 shows the computed values of $\bar{I}_B$, monthly averaged normalized diurnal variabilities, for 35 stations in Israel during the months of July and January 1978. The values in parentheses represent the January numbers. The complex topography consists of: (from west to east) the Eastern Mediterranean seashore; the Central Mountain region; the Negev and Sinai; the Jordan and Arava valleys, which contain the Sea of Galilee, the Dead Sea and the Red Sea; and the Golan Heights to the northeast.

The diurnal relative gustiness $\bar{I}_B$ is in general higher in summer except in the Central Mountain region and the southern gulfs; see, e.g., Jerusalem, where in winter $\bar{I}_B$ = 0.47 as compared to 0.34 in summer. The larger diurnal relative gustiness in the summer is particularly noticeable in the Jordan Valley. All of the values are larger than the typical turbulence intensities, 0.15–0.25 as discussed before, and reflect the vigorous diurnal wind variability.

Comparison of the calculated values of $\bar{I}_B$ from Fig. 1 with the contribution to $\bar{I}_B$ by the stratification changes estimated by ~0.11 shows that the latter contributes less than ~30% in nearly all stations. That means that the major contribution to the diurnal wind variability in the Israel region is due to the local mesoscale forcing, while the effect of the diurnal changes in the thermal stratification is secondary. Of course, there is also a contribution to the diurnal variability due to diurnal changes of the large-scale pressure-gradient, but this effect is minimized by the averaging of I_B over a period of a month. For comparison, the values of I_B at O'Neill, Nebraska, based on Lettau and Davidson (1957) data were also calculated for six of the observation periods; see Table 1. The fourth observational period didn't cover a full diurnal cycle. For the rest of the periods we have used the first 24 hours from the data. The calculations are based upon the hourly data, but we have also checked I_B values from quarter-hourly wind data and the differences were negligible. Notice that the I_B values from Lettau and Davidson's data contain the large-scale contribution as well, because no monthly average, such as the one we calculated, could be performed. Also, the mesoscale contribution is supposed to be relatively small at the quite homogenous O'Neill site. Therefore, the third and fifth observational periods, which were characterized by low large-scale variability [see Lettau and Davidson (1957, pp. 447, 489)], exhibit the lowest I_B values. The first and seventh observational periods, for instance, which were subject to considerable large-scale diurnal changes, indicate higher values. As already mentioned, this large-scale contribution, which is occasionally strong in our study, is mostly reduced[2] by the calculation of $\bar{I}_B$ rather than I_B. It is interesting to note that the lower I_B values at O'Neill (0.16) are quite close to the theoretical estimation of 0.11 over horizontally homogenous terrain, based upon the Wipperman *et al.* (1973) model.

In Fig. 2 the values of the "normalized interdiurnal variability" I_A are presented for the same time and locations. From Figs. 1 and 2 one can clearly see that in summer the interdiurnal variability I_A is generally smaller than the averaged diurnal variability $\bar{I}_B$; i.e., $I_A \leq \bar{I}_B$. In winter this relation is, in general, reversed;

TABLE 1. The values of the "diurnal relative gustiness" at O'Neill, Nebraska based on Lettau and Davidson (1957) data for six observational periods.

	Observing period						
	1	2	3	4	5	6	7
$I_B = \sigma_v/\bar{V}$	0.34	0.23	0.16	—	0.16	0.16	0.27

i.e., $I_A \geq \bar{I}_B$. If a tilde denotes an average over all the stations, then

$$\left.\begin{aligned}\tilde{I}_A &= 0.21 \pm 0.09 \quad (0.4 \pm 0.10)\\ \tilde{I}_B &= 0.44 \pm 0.09 \quad (0.39 \pm 0.06)\end{aligned}\right\},$$

where the values in parentheses are for the winter results. In the calculation of these averages, the values for five additional stations (not drawn) were also considered. It is interesting to note the particularly low interdiurnal variability during the summer in the Jordan and Arava valleys and also in the southern deserts. The overall geographical variability in I_A is small: only five stations have an I_A value larger than 0.3 and only four have values below 0.12. This low scattering of the "gustiness" values is noticeable for I_B also. Moreover, the stations at the same geographical–climatological regions exhibit quite similar values of I_A and I_B. For instance, notice the summer values of $\bar{I}_B$ for the five stations over the Golan Heights.

The fact that the interdiurnal normalized wind variability I_A exhibits quite low values in Israel (an average of I_A = 0.21) is associated with the well-known persistent synoptic conditions of the Eastern Mediterranean during summer. The higher-level subtropical ridge extending from North Africa to Israel prevails during the summer. A nearly permanent inversion at an average altitude of about 1000 m prevents any rain clouds from developing, and the shallow monsoonal trough supports the weak northwesterlies at the surface. The diurnal wind variability is due to local mesoscale forcing as well as land–sea breezes and mountain–valley winds which are superimposed on the aforementioned quite steady large-scale pressure gradient. The interdiurnal wind variability is primarily due to small oscillations in the position of the monsoonal trough and the subtropical ridge.

The aforementioned synoptic pattern explains why I_A is relatively small during summer. In some stations I_A drops below 0.15; for example on the northern coast of the Dead Sea I_A is 0.13 and in northern Negev it is 0.10 in two stations and 0.12 in Ktziot. Such values are considered small even when compared to the typical turbulence intensities, which vary between 0.1–0.25; see, e.g., Sutton (1953), Swanson and Cramer (1965). Therefore, when these values are compared to the diurnal wind variabilities expressed by $\bar{I}_B$ which are much larger, one finds that the diurnal wind variability dominates during the sum-

[2] Most of the days that have been studied are characteristic of very small synoptic (large-scale) variability. Therefore averaging I_B over a month will tend to bring the large-scale contribution to the relatively small monthly average.

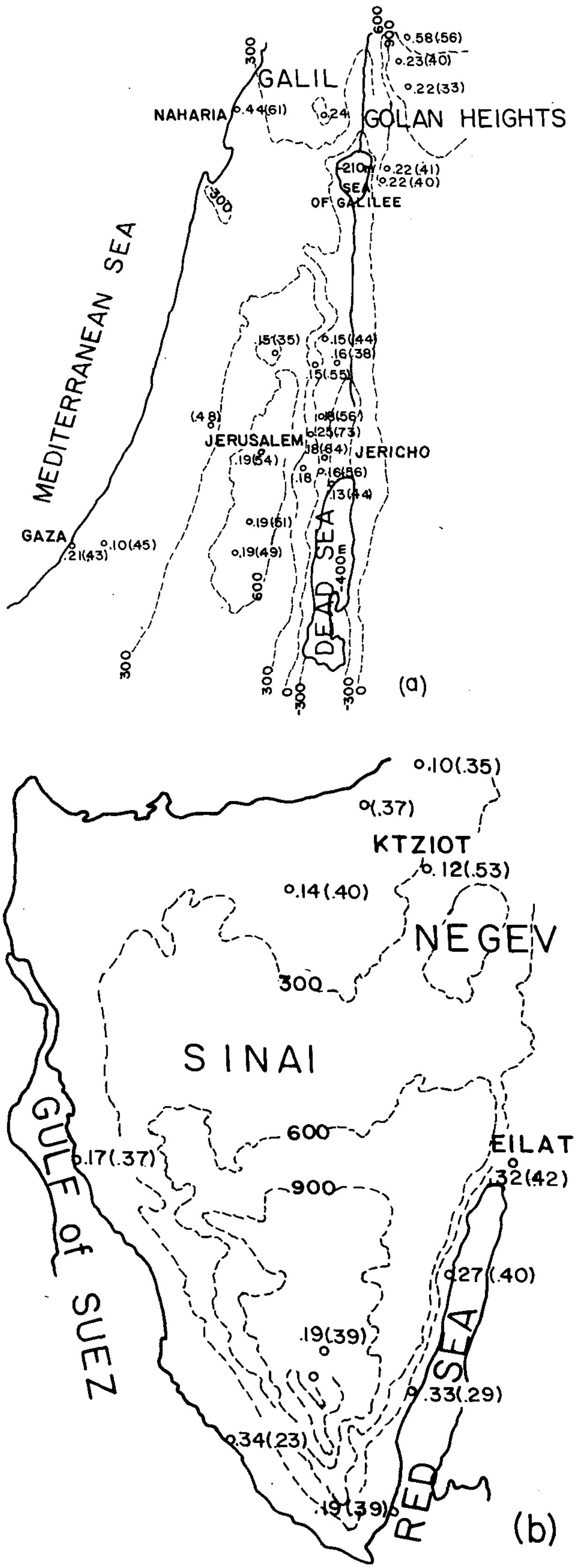

FIG. 2. As in Fig. 1 but for the normalized interdiurnal variability I_A.

mer. This comparison is performed by the calculation of $\alpha = \bar{I}_B/I_A$ in the following section.

4. Index for mesoscale wind activity—observations

The calculated values of the index for the diurnal to interdiurnal relative variabilities α [see Eq. (6)] are depicted in Fig. 3. Nearly all stations show summer values greater than one and winter values less than one. This illustrates that the area under investigation is dominated during summer by the diurnal rather than interdiurnal processes in producing wind variability, whereas in winter the situation is just the opposite. This means that in winter the diurnal boundary layer forcing plays a less significant role than do the migrating troughs and ridges, which are associated with the motion of the midlatitude synoptic systems. Clearly, the latter effect becomes secondary during summer in this region. Notice the relatively high α index in the Jordan Valley and the Negev during summer; it exceeds 2–3 or even 4–5. These regions are characterized by high diurnal wind variabilities relative to interdiurnal values. In the Jordan Valley the high α-values are due mostly to the sheltering of the mountains on both sides and the channeling effect, while in the Negev and Sinai deserts the main reason for the high α-values is probably the dominance of the local pressure gradient forces induced by the vigorous diurnal differential heating over the desert boundary layer. The average α-value $\tilde{\alpha}$ ($\tilde{\alpha} = \tilde{I}_B/\tilde{I}_A$) for all the stations that were measured in this region is 2.09 (0.84) in summer (winter).

There are few stations in the southern deserts and the Gulfs of Eilat and Suez that show $\alpha > 1$ even during winter. This reflects the fact that the influence of the migrating winter cyclones is much weaker in the southern region. The tracks of the winter cyclones generally miss the southern region and thus the annual precipitation there, for instance, is much lower than in the northern or central parts of Israel; see, e.g., Katsnelson (1964). Therefore, differential heating and mesoscale forcing play a significant role in the southern regions even during the winter.

We suggest that a regional map of α-indices, like that shown in Fig. 3, indicates areas in which mesoscale models will probably be highly successful even though they ignore variability due to the large-scale pressure gradients. Such examples in the Israel area were given by Alpert *et al.* (1982) in the prediction of the diurnal wind in the Sea of Galilee (Lake Kinneret) area or by Segal *et al.* (1983) in the calculation of the diurnal wind changes near the Dead Sea. According to our suggestion, Fig. 3 indicates that the whole area is highly suitable for mesoscale modeling during summer. During winter, however, mesoscale models might be useful only in the southern part of the region, unless the effect of the large-scale pressure-gradient variability is incorporated into the model. As previously discussed, many of the mesoscale models neglect the latter effect.

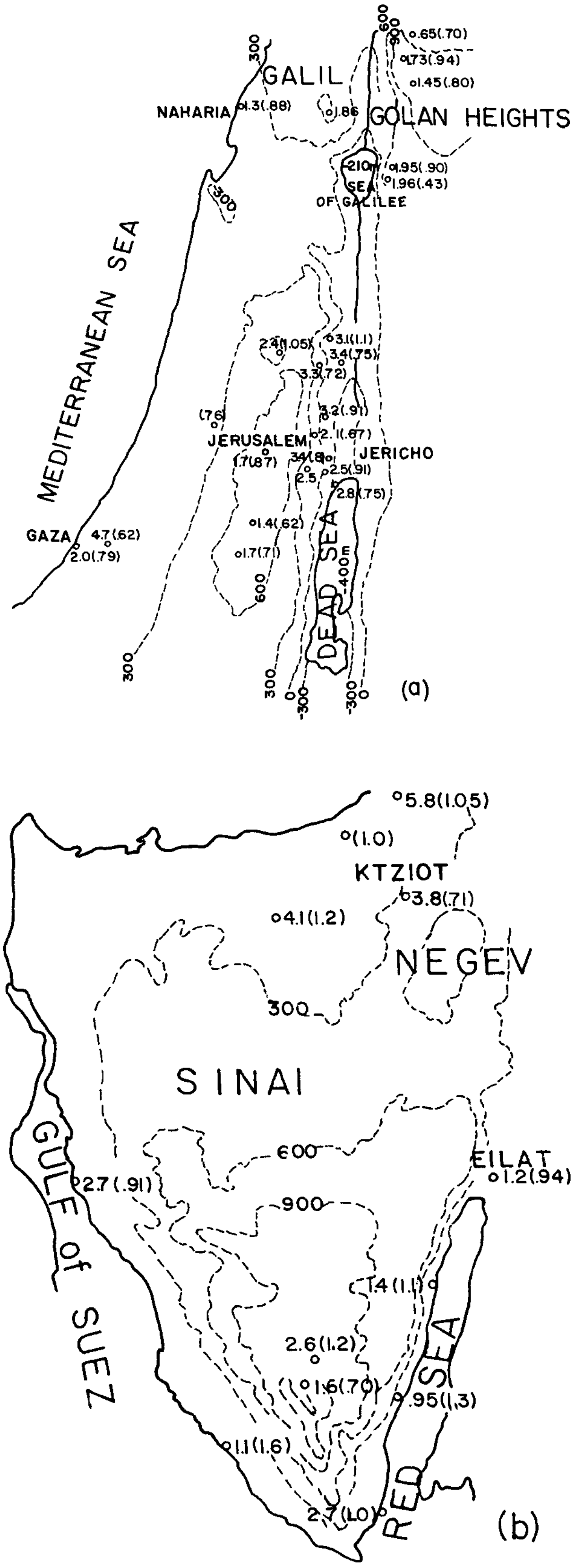

FIG. 3. As in Fig. 1 but for $\alpha = \bar{I}_B/I_A$, the index for mesoscale activity.

5. Representativeness of the average wind hodograph

An interesting way of presenting diurnal wind measurements, as discussed by many investigators, is the hodograph. In this section we would like to address the following question: Does the average hodograph represent faithfully the actual wind measurements? The answer to this question depends upon dominance of the diurnal wind variability relative to the interdiurnal changes. Therefore we will suggest that the α index serve also as a measure of the extent that any hodograph is a faithful representation of the actual wind observations. As an example, Figs. 4–7 present eight averaged hodographs for four stations in the Israel region during July and January 1978. The stations are situated in different climatic–geographic regions and their locations are indicated in Figs. 1–3. The α index for each of the hodographs is also indicated in Fig. 3.

The wind hodographs of Jericho during summer and winter (Figs. 4a, b) illustrate the idea that an elliptical shape (elliptical because such a shape is predicted by, for example, the linear theory for the sea breeze—see Haurwitz, 1947) of the observed wind hodograph is commonly associated with a high value of diurnal to interdiurnal wind variability. The very clear and simple summer hodograph (Fig. 4a) shows a high α index, whereas the complex winter shape is associated with a lower value. This feature is common to many of the hodographs observed but there are exceptions. The winter hodograph at Ktziot has a much clearer shape than its summer counterpart (Figs. 5a, b), although the latter has a higher index. A similar argument is valid for the Eilat hodographs (Figs. 6a, b). In other words, a simple hodograph does not guarantee a stronger diurnal wind variability relative to the interdiurnal one. A fourth example is given by the Gaza hodographs (Figs. 7a, b). Both July and January hodographs are clear and simple ellipses, although the α indices differ appreciably. But the α indices indicate to what extent the average wind hodograph represents the actual wind on a particular day. For example, in Gaza the averaged north-northwesterly wind at 1800 LST during summer is much more representative of the actual wind at that hour on a particular summer day than the average (nearly) westerly wind at the same hour for winter.

6. Discussion and conclusions

Lumley and Panofsky (1964) indicate that the "relative gustiness" of the wind, which represents the variance of the longitudinal velocity, varies in neutral air with the terrain [according to the constant C in Eq. (2)], along with the logarithmic dependency upon z/z_0, z_0 being roughness length. In this study we extend the definition of the relative gustiness to longer averaging time intervals in order to define the diurnal and interdiurnal wind variabilities in similar "gustiness" terms. It is suggested that the α ratio between

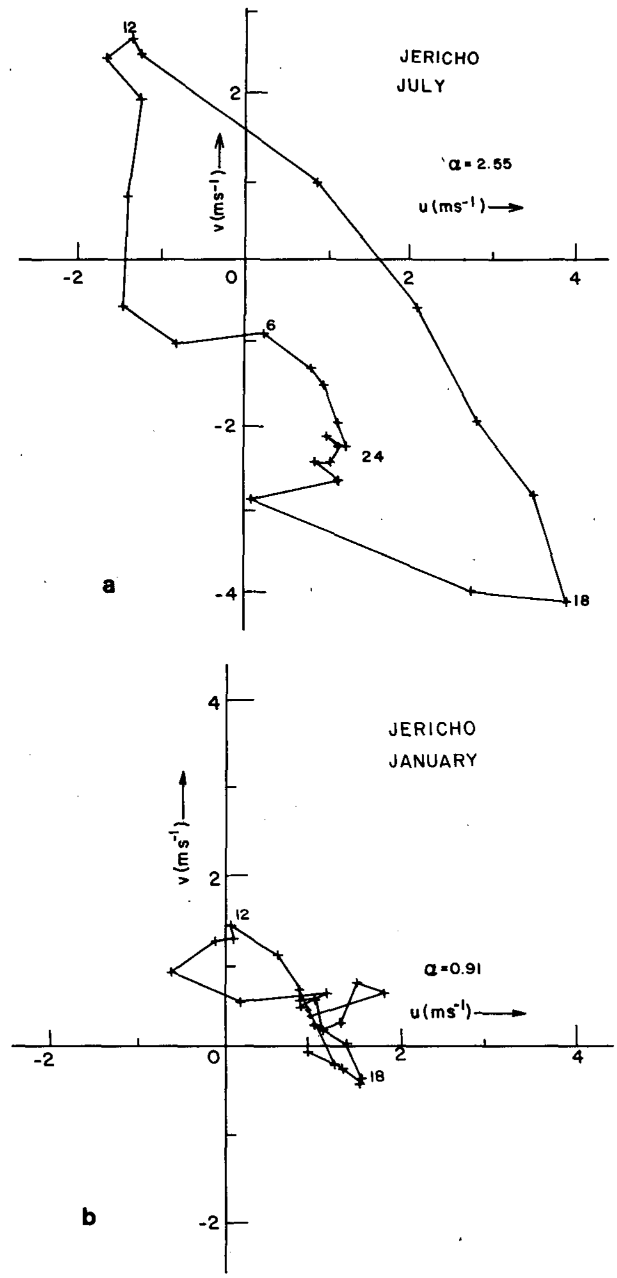

FIG. 4. The wind hodograph of Jericho, Jordan Valley (a) for July 1978 and (b) for January 1978.

the diurnal and the interdiurnal wind variabilities, which are defined in terms of this extended relative gustiness, may serve as a useful index in mesoscale studies. The α index is calculated for more than 30 stations in the Israel area during summer and winter and is found to be a good indicator for the following features:

1) A region with high α index, i.e., $\alpha > 1$, is characterized by a diurnal wind variability that is dominant over the interdiurnal one. The higher the α-value is, the higher the chance for a successful prediction of the wind by mesoscale models that ignore the transient effects of the large-scale pressure gradient variations. In such cases, the diurnal processes which are forced by boundary layer differential heating play the most significant role in producing the diurnal wind variability. If the α-index is smaller than 1, one cannot, in general, neglect the contribution of the

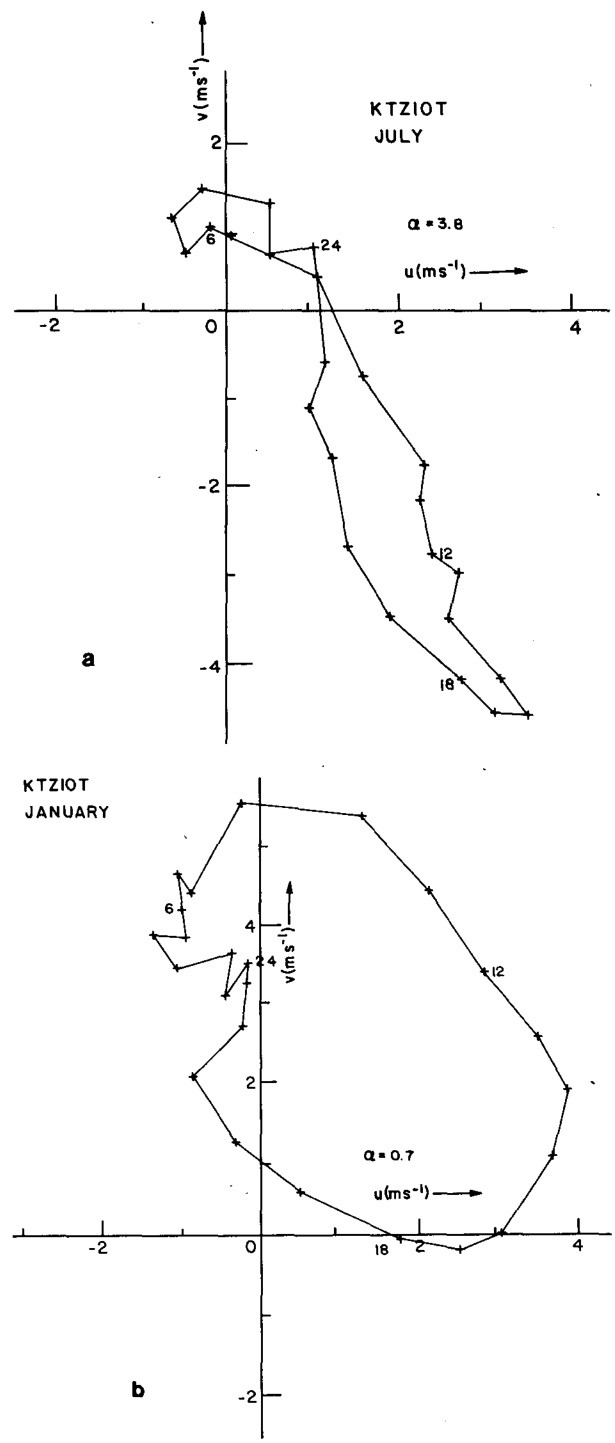

FIG. 5. As in Fig. 4 but for Ktziot, Negev Desert.

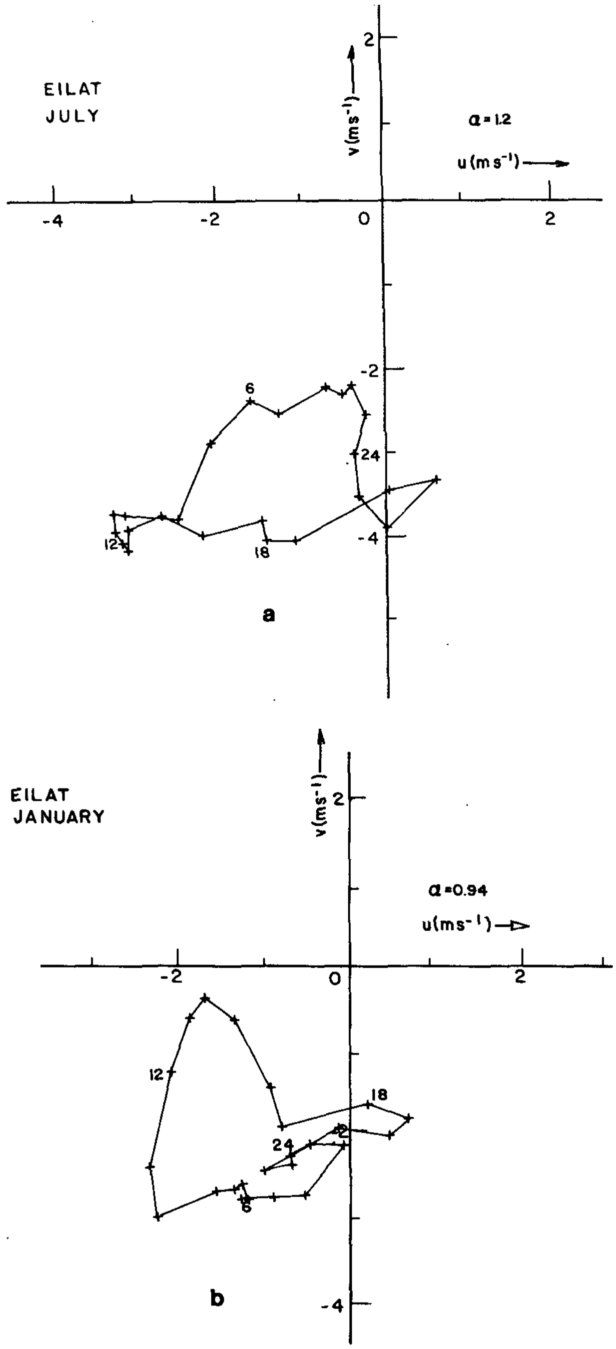

FIG. 6. As in Fig. 4 but for Eilat, southern Arava Valley.

large-scale variations in producing wind variability. In such climatic regions, wind prediction by mesoscale models would probably not be as successful unless large-scale variations are incorporated in the model. Most of today's mesoscale models assume a constant large-scale pressure gradient.

2) An interesting way of presenting wind measurements, as discussed by many investigators, is the hodograph. The α index measures the degree to which the average wind hodograph is a faithful representation of the actual diurnal wind changes on a particular day. An α index greater than 1 suggests a more faithful representation of the actual wind by the average hodograph. In the case $\alpha < 1$, the hodograph does not represent the actual wind faithfully, since the interdiurnal wind variability is dominant over the diurnal one.

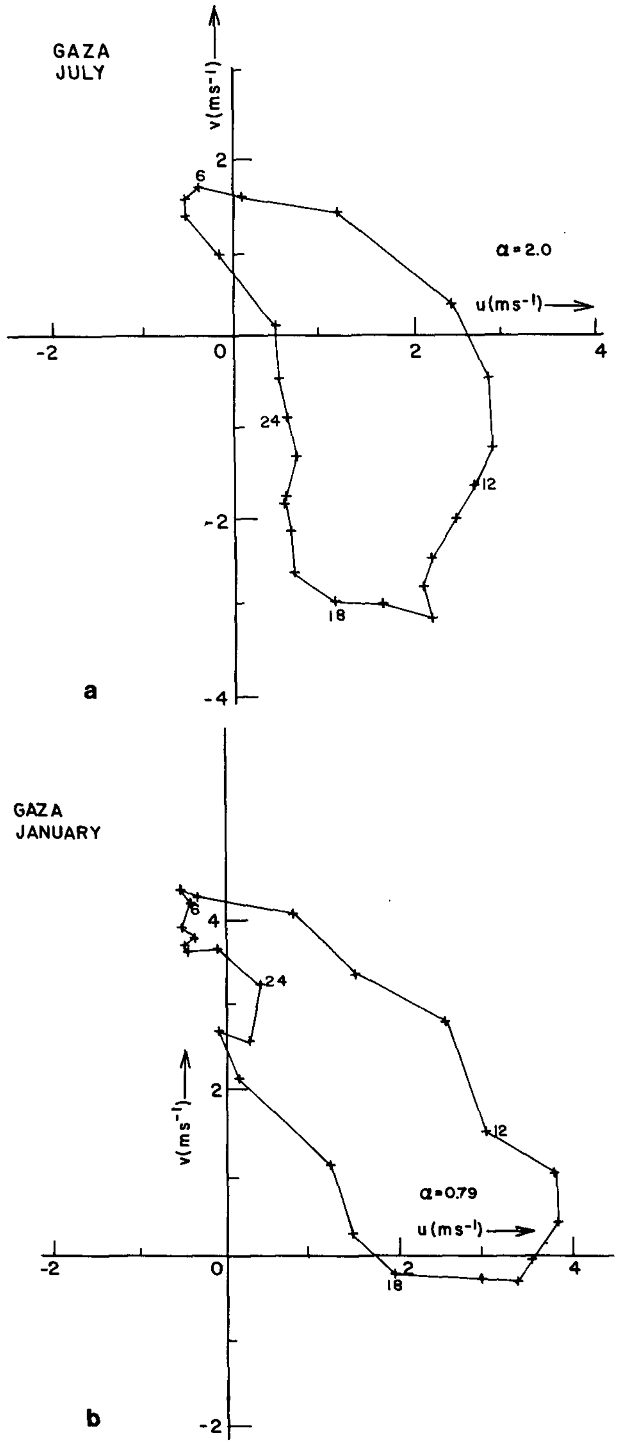

FIG. 7. As in Fig. 4 but for Gaza, southern Mediterranean coast.

It should be mentioned, however, that the identification of the α index with mesoscale activity is not rigorous since we have ignored the microscale contribution to the wind variability by the diurnal stratification changes over homogenous terrain. The latter contribution was estimated to be 0.11–0.16. A pure mesoscale index should subtract this contribution from the diurnal wind variability $\bar{I}_B$. The effect of such a correction would be to diminish slightly the α-values during summer, since the $\bar{I}_B$-values are quite high, and to considerably decrease the α-values during winter, when the $\bar{I}_B$-values are small and are comparable in their magnitude both to the interdiurnal contribution I_A and to the microscale contribution. The overall result would enlarge the significant differences between the α indices during summer and winter and the α will then better serve as an index for mesoscale activity. This was not done since the value of the microscale contribution is uncertain, and, according to the aforementioned estimation, our major conclusions are not influenced by this correction.

Also, it should be stated (as correctly pointed out by one of the reviewers) that there are other synoptic-scale variables, besides the wind, whose variability can affect a mesoscale simulation. For example, in simulating precipitation an index based upon the moisture variabilities might be even more appropriate. However, we believe that for dry atmospheric circulations α is a very convenient (easy to calculate) and most appropriate index for mesoscale activity.

Based upon the relatively great experience of mesoscale modeling that has been gained in the Israel region, the α index is found to reveal the areas with the highest potential for mesoscale studies. Nearly all stations show $\alpha > 1$ during summer, which indicates high mesoscale activity. In some regions such as the Jordan Valley, Arava and the Negev and Sinai deserts the α index attains particularly high values ($\alpha \approx$ 2–4). This reflects both the vigorous mesoscale forcing due to the channeling effect and/or strong differential heating and the reduced effect of the migrating cyclones during the Israeli summer. In winter the situation is almost the opposite; this is illustrated by α values which are less than one in most of the climatic–geographical regions. It follows that the α index may serve as a powerful though simple tool to reveal areas that are appropriate for mesoscale modeling even though they ignore the effects of the large-scale variations upon the evolving mesoscale wind fields. Unfortunately, most modern mesoscale models that simulate evolving boundary layer forced circulations have neglected this effect without any good physical justification. It should be stated, however, that it is important to be able to do mesoscale modeling without having to include large-scale effects. This is so because the study of *pure* mesoscale circulations by comparison with observations must rely on situations where we can assure that the transient large-scale effect is realistically unimportant, i.e., α exceeds 1 considerably. We believe that most of the successful mesoscale simulations presented in the literature over the last two decades were performed in regions where the α indices were quite large. It is suspected that most of the midlatitude regions possess α indices smaller than 1 during most of the year. Therefore, successful mesoscale wind simulation requires the incorporation of the variable large-scale pressure gradients.

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