NUMERICAL STUDY OF
WIND ENERGY CHARACTERISTICS
OVER HETEROGENEOUS TERRAIN –
CENTRAL ISRAEL CASE STUDY

MORDECAI SEGAL
Department of Environmental Sciences, University of Virginia,
Charlottesville, Virginia 22903, U.S.A.

YTZHAQ MAHRER
Department of Soil and Water Sciences, The Hebrew University of Jerusalem, Israel

and

ROGER A. PIELKE
Department of Environmental Sciences, University of Virginia,
Charlottesville, Virginia 22903, U.S.A.

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Abstract. A numerical mesoscale meteorological model has been applied over the heterogeneous terrain of central Israel in order to study wind energy characteristics of three typical synoptic situations. The supportive nature of this method for observationally oriented wind energy studies has been emphasized. Mesoscale forcing effects on the availability of wind energy and on the exponent, $p$, in the vertical wind power law are evaluated.

1. Introduction

In recent years there has been increased interest in the search for new energy resources, including wind power (WP). The selection of optimal WP sites is an involved interdisciplinary procedure; e.g., see the discussion by Hewson (1975). Assessments of WP sites are based on various statistical processings of wind speeds preferably at several levels above the ground; see, e.g., Justus et al. (1976), Hennessey (1977). Wind speed data are obtained from surface observations, mostly from synoptic stations, or, occasionally, from special networks such as those designed for air quality or agrometeorology studies. However, two major shortcomings are characteristic of such data sources:

- Generally, the spatial density and siting of these networks are inadequate to meet the needs of WP surveys.
- Anemometer heights in these networks are usually in the range of 5–15 m, while wind speed information at higher elevations is required for large wind-power generator units. Hence, profile extrapolation procedures must be utilized. Usually, these extrapolations are based on power-law formulae, although the accuracy in the determination of the exponent $p$ in such relationships cannot be examined at sites where higher level wind data are absent.

These shortcomings are enhanced in regions where meso-scale flows occur.

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In this paper, a numerical mesoscale model is suggested in order to acquire better spatial and temporal resolution of WP characteristics. In addition, such model simulation results can supply guidelines for WP measurement strategies. This approach has also been proposed by Garstang et al. (1980a) and was the basis for the WIVEX observation field program in January 1980 over the Chesapeake Bay (Garstang et al., 1980b). The use of a mesoscale model is especially advantageous in regions where long periods with a high degree of steadiness of the daily flow patterns are common. In those locations, simulated WP patterns should closely correspond to patterns obtained by statistical methods based on observed data.

Generally, a synoptic classification is needed in order to develop a climatology of weather pattern type for input to a mesoscale model. Lindsey (1980) reports on such a methodology for the Gulf and Atlantic coasts of the United States; the method has been used as input to the University of Virginia Mesoscale Numerical Model (UVMM) for WP evaluations (Snow, 1981).

In the current study, the UVMM has been applied to central Israel, to simulate WP characteristics associated with the following situations:
- Summer Day
- Adveective Sharav
- Winter Day
where, for practical WP considerations, the first case is of the most importance. Unlike the aforementioned UVMM WP simulations, however, in the present study the coastal region includes mountainous terrain, resulting in a further enhancement of mesoscale effects on the WP patterns.

Results in this paper include evaluations of the spatial and temporal distribution of the WP and the exponent, $p$, of the wind power-law extrapolation formula as computed by the model. The effect of mesoscale forcing on these two characteristics is discussed along with illustrations of the flow and thermal stratification patterns within the lower atmospheric layer.

2. Description of Simulated Terrain

Figure 1 illustrates schematically the simulated terrain. To the west there is a narrow strip of the Mediterranean Sea and the adjacent coastal plain of central Israel. Inland, the topography increases gradually to a typical elevation of about 800 m at the top of the inland mountain ridge (consisting of the Judean mountains in its southern section and the adjacent Shamarian mountains to the north). Farther east is the Jordan Rift Valley (JRV), 300 m below mean sea level (MSL). The Dead Sea ($-400$ m MSL and 85 km in length) is located in the JRV, with a typical width of about 15 km. To the east, the Amman plateau has an elevation of about 700 m.

The coastal plain is more highly populated (and, therefore, more built up) than the rural western slopes and the arid eastern escarpment. Because of this variation in ground characteristic, surface roughness decreases eastward across central Israel. Generally, the topography is closely symmetric along a line that is almost north-south.
Fig. 1. Topographical illustration of the simulated domain; crossed line represents the inland mountain ridge, dashed-dotted contours are for terrain height below mean sea level. Contour values are given in meters. The heavy dark line indicates the model cross-section.

3. Model Aspects

The UVMM, which provides simulations of terrain-induced mesoscale systems, was developed originally by Pielke (1974) with most of the recent modifications reported by Mahrer and Pielke (1978). The model is hydrostatic, consisting of the equations of motion, heat, moisture and continuity in a terrain-following coordinate system. Verification studies of the model in recent years have included, for example, those of Pielke and Mahrer (1978), Mahrer and Segal (1979a) and Segal and Pielke (1981). In the present paper only a brief description of the model is given.

3.1. Planetary Boundary Layer (PBL)

The calculations of the surface fluxes of momentum, heat and moisture are based on the work of Businger (1973). The exchange coefficients in the PBL above the surface layer utilize O’Brien’s (1970) functional form. The PBL top is predicted as a function of location according to Deardorff’s (1974) prognostic equation. (Test runs in the current study indicated that the incorporation of Smeda’s (1979) PBL height prediction equation for the stably stratified surface layer resulted in only a slight effect on the simulated fields). The roughness parameter over the land is prescribed, while over the water it is calculated according to Clarke (1970).
3.2 Surface Heat Balance

The temperature at the soil-air interface is calculated using a heat balance equation, which includes solar radiation, incoming atmospheric long-wave radiation, latent, sensible and soil heat fluxes and the outgoing surface long-wave radiation.

3.3 Radiation

The changes of air temperature due to short-wave and long-wave radiative flux divergence are parameterized following methods of Atwater and Brown (1974). Heating of the atmosphere by short-wave radiation is confined to water vapor, while carbon dioxide and water vapor are considered in the long-wave radiation heating/cooling algorithm.

3.4 Initial and Boundary Conditions

The initial atmospheric data are wind velocity, temperature and humidity. The initial wind velocity above the initial specified PBL is geostrophically balanced; within the PBL

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Fig. 2. Vertical profiles of the synoptic wind velocities used as initial flow conditions for the three simulated cases.
it is determined by assuming a balance of the shear stress, the Coriolis and the pressure gradient forces.

Zero gradient conditions are assumed at the model lateral boundaries.

3.5 General Aspects

In the present study, the two-dimensional version of the model has been adopted, since, as discussed in the previous section, the terrain is reasonably two-dimensional throughout most of the domain (excluding the bottom of the JRV, which consists of water and land terrains). The geographical location of the model cross-section is indicated by a dark line in Figure 1. (It implies that, at the bottom of the JRV, the model-predicted fields represent the region with water.)

The two first simulations consist of 14 vertical levels (as shown in Figure 2). A test simulation of the summer case, with a much higher vertical level resolution (see Figure 6 for the heights used in the lower 100 m of the atmosphere), revealed that the WP characteristics can be well resolved with the adopted less dense vertical resolution (14 levels) as illustrated in Figure 2. Due to the strong synoptic flow associated with the winter day case, seven vertical levels have been added for this simulation, elevating the model top to 21 km. This modification, for the winter case, was introduced to minimize possible erroneous downward reflection of mountain waves from the model top to its surface boundary, as discussed by Mahrer and Pielke (1978), Anthes and Warner (1978) and Kessler and Pielke (1981).

The model horizontal grid interval is 5 km and the integration time step is 90 s. The roughness parameter, $Z_0$, is 4 cm over the land except at the coastal plain, where it has been modified to 10, 25, 20, and 15 cm at distances of 0, 5, 10, and 15 km onshore, respectively. There are some asymmetric areas within a strip of about 10 km onshore, due to urbanization; hence, higher $Z_0$ values might be appropriate here. An accurate simulation of the airflow over these urbanized regions would require a three-dimensional model but, since this area is one of reduced WP for the most persistent flow (as will be shown), the 2-D model is practically appropriate for a regional evaluation.

Model simulations were carried out for 28 h, beginning at 2000 LST (after sunset) after a 6 h period of dynamical initialization. (The dynamical initialization involves the integration of the model equations, in the absence of heating sources, in order to balance the initially imposed mass and motion fields.) The initial vertical profiles of the wind velocity for each of the three simulations are given in Figure 2. Clear skies are assumed for these simulations, although such conditions do not always occur in the last two cases.

4. WP Computations

WP has been computed each time step at selected model grid points according to the formula

$$E = \frac{\rho}{2} V^3.$$  

(1)
In (1), $V$ is the wind speed and $\rho$ is the air density (variations of $\rho$ due to topographical changes and also those due to different computation levels, were included). For better representation of the WP, we have utilized hourly averages (as is commonly done in WP surveys). The hourly-averaged WP at hour $T$, was computed from

$$ WP_T = \int_{T-1}^{T} E \, dT. $$

(In the integral discretization, $dT$ corresponds to the simulation time step given in a fraction of an hour.)

Additionally, a daily averaged distribution of the WP was computed according to

$$ WP_D = \int_{D-1}^{D} E \, dD. $$

where the integral is averaged over a day rather than an hour. The averaging begins at 2300 LST in order to minimize the influence of the initialization.

5. Wind Power-Law Exponent ($p$) Computations

A power law is frequently used to estimate wind speed at upper levels from surface data, i.e.,

$$ \frac{V}{V_r} = \left( \frac{Z}{Z_r} \right)^p $$

where $V_r$ is the observed wind speed at a reference height, $Z_r$, $p$ is a constant and $V$ is the extrapolated wind speed at height $Z$. Usually, $p$ is prescribed as dependent on the stability of the atmosphere near the surface (e.g., using the Pasquill stability classification, or its equivalent in terms of vertical temperature gradient).

Touma (1977) has examined published wind data for many locations indicating that $p = \frac{1}{3}$ is appropriate for the neutral atmosphere, while higher values of $p$ (reaching $\sim 0.5$) are needed when the surface layer is stably stratified. Peterson and Hennessy (1978) however, claim that whenever the terrain is not too complex, stable atmospheric conditions are not associated with practical WP and suggest that $p = \frac{1}{3}$ is adequate for realistic but conservative estimates of the wind power at different heights (except at extremely rough sites where the estimates may be conservative).

An accurate evaluation of $p$ within the surface layer can be derived from (4) using surface-layer similarity theory, expressed as:

$$ p = \ln \left( \frac{\ln \left( \frac{Z}{Z_0} \right) - \psi \left( \frac{Z}{L} \right)}{\ln \left( \frac{Z_r}{Z_0} \right) - \psi \left( \frac{Z_r}{L} \right)} \right) / \ln \left( \frac{Z_r}{Z_r} \right). $$

$$ (5) $$
In this relation, the stability correction term, $\psi$, to the nondimensional wind function is that given by Businger (1973), where $L$ is the Monin–Obukhov length. Sedefian (1980) has used an expression similar to (5) for constructing a useful nomogram evaluating $p$ as a function $Z_0$ and $Z/L$. He suggests, in addition, a way of estimating $L$ based on ground-level observed data as discussed in Golder (1972); a method which allows the simple practical use of (5). In the derivation of (5), however, it is assumed that mesoscale horizontal gradients of wind velocity and pressure are negligible within the surface layer. In inhomogenous terrain, as in the present study, this assumption is not necessarily satisfied, while, in addition, WP levels of interest may exceed the surface layer depth. Then, the only way of estimating $p$ is from (4) when wind speed at both the reference level and the level of interest are known.

Using (4) and the model hourly and daily wind speed predictions, we have derived hourly and daily averaged $p$ values.

6. Results

For each simulated case, the results consist of hourly averaged WP distributions at 5, 15, 50, and 100 m, and the hourly averaged $p$ values at 15, 50, and 100 m, using a height of 5 m as the reference level. The wind velocity and relative vertical temperature change distributions, at the end of the averaged hour, are also illustrated. Additionally, daily averaged distributions for the WP and $p$ are included.

The following remarks are useful in interpreting the results:

(a) A common feature in the three simulations is the substantial influence of topography on the synoptic flow, viz., the reduction of the flow intensity along the windward slopes and its intensification on the top or the leeward side.

(b) As the experiments have been carried out for clear sky conditions, the most pronounced daytime thermally induced flows occur in the summer, when the solar insolation is greatest.

(c) Production of power by wind generators is limited by lower and upper wind speed thresholds (depending on the machine type). It is common practice in detailed seasonal or annual WP evaluations to restrict wind speeds to the range resulting in effective WP, e.g., Justus et al. (1976), Hennessey (1977), Baker et al. (1978). In the present study, we have not considered the economical and engineering aspects of WP. Rather, the total available WP has been computed, with no cut-in and cut-out limitations considered.

6.1 Summer day

During the summer (from about the period of mid-June to mid-September), the synoptic surface conditions over the region are characterized by a permanent trough resulting from the monsoon pressure depression located over the Himalayan mountains (Figure 3a). This produces a persistent westerly-northwesterly gradient wind estimated by Doron and Neumann (1977) to be about 3 m s$^{-1}$. Coupled with sea and land breezes and mountain-induced thermal flows, this leads to a well-marked daily surface wind cycle (Skibin and
Fig. 3. Schematic illustration of surface synoptic systems typical of: (a) summer day-July averaged pattern, (Based on Jaffe, 1976); (b) and (c) advective Sharav (based on Levi, 1967); (d) winter day (based on Atlas of Israel, 1963).

Hod, 1979). Although changes in the flow patterns are expected during this season (due to changes in mesoscale thermal forcing intensity as well as in the location of the surface trough), the impact can be considered of secondary importance. Consequently, the
computed WP patterns for the summer can be regarded as characteristic for a substantial portion of the year. Following Mahrer and Segal's (1979a) verification case study, July 15, 1977 has been chosen to represent typical summer conditions.

At 0500 LST (sunrise hour) the WP patterns represent the nocturnal flow at the end of the night (Figure 4). Along the coastal plain and the western slopes of the inland

Fig. 4. Several selected hourly and daily averaged WP distributions as a function of distance from the coastline at 5, 15, 50, and 100 m levels (upper portion of each figure) and the associated wind velocities at the end of the hour (lower portion of each hourly averaged figure). The horizontal components of the wind velocity vectors in the Figure reflect east-west flow; the vertical component is for the south-north flow. The topography is illustrated by a dotted line; the water section of the Mediterranean Sea and the Dead Sea are indicated by heavy dark lines. Summer day.
mountain ridge, the nocturnal land breeze and the downslope mountain flow oppose the prevailing westerly flow, resulting in poor WP availability. On the other hand, the dynamic intensification of the synoptic flow to the lee of the mountains is enhanced by downslope nocturnal drainage, causing a WP increase there. As the Mediterranean Sea breeze begins penetrating inland during the daylight hours (1100 LST), its unidirectional coupling with the synoptic flow in the coastal area has relatively little effect on WP availability. Thermally-induced upslope flow, in addition to the deepening of the planetary boundary layer within the vertically sheared synoptic flow, however, increases WP along the western slopes and at the mountain top. Along lee slopes, thermally induced upslope flow opposes the synoptic flow, resulting in a reduction of WP from the nocturnal values. The Dead Sea lake breeze also induces some WP, although it is confined to the lake area. (In our simulations, the model prediction along the JRV is limited to its Dead Sea segment. North of the Dead Sea, upslope flows are due to mountain thermal effects only and, thus, are expected to be weaker than those along the Dead Sea segment.) At 1800 LST, the Mediterranean sea breeze sweeps down the eastern slopes, creating a secondary eastward moving WP peak. This downslope flow eliminates the local Dead Sea lake breeze and its associated WP. The daily averaged pattern indicates highest WP availability along the eastern slopes (the section 50 to 70 km inland), mostly at ridge top.

Baker et al. (1978) suggested a cut-in wind speed of 4.5 m s\(^{-1}\) and cut-out wind speed of 27 m s\(^{-1}\) for effective WP. Additionally, they have classified the highest seasonally averaged effective WP when the effective WP exceeds 400 W m\(^{-2}\) and power-producing winds occur for more than 50% of the season. The lowest rate is when the effective WP is less than 200 W m\(^{-2}\). It is evident that wind speeds are typically above 4.5 m s\(^{-1}\) at the location of the maximum daily averaged WP. Adopting the classification of Baker et al. indicates the availability of adequate effective WP there.

Based on the high degree of persistence of the diurnal surface flow cycle over Israel during the summer, Doron (1979) has performed a quantitative objective analysis of the averaged July surface flow. Scarcity of available data at the mountain ridge and elevated eastern slopes in his analysis excluded the possibility of adequate comparisons between model-predicted and observed wind speeds for this potential location of effective WP. However, model surface predictions are generally supported by his objectively analyzed flow patterns. Additionally, it is worth noting the reasonable agreement between the observed winds pattern reported by Bitan (1977) at Mizpe-Shalem Fields (located on the western Dead Sea shore line of the model cross-section) and the model predicted winds.

The inland distribution of \(p\) during the nocturnal thermally stable period (0500 LST) shown in Figure 5 indicates that along most of the coastal plain and western slopes, \(p\) values, computed at 15 m, are much larger than those at the 50 and 100 m levels. These differences in \(p\) values reflect the complex nature of the flow when a shallow nocturnal layer, with an easterly component near the ground, diminishes gradually aloft as the westerly synoptic flow becomes dominant. On the other hand, along most of the eastern slopes, where both flows are in the same direction, \(p\) values are similar for all levels.

Inland, during the daylight hours (1100 LST and 1800 LST) when thermal stratifi-
Fig. 5. $p$ (exponent in wind-power extrapolation law) distribution as function of distance from the coastline for the heights of 15, 50, and 100 m; based on 5 m as the reference height, at several selected hours (upper portion of each figure) and the daily averaged $p$ values. In the lower part of each hourly averaged $p$ figure, the relative change of temperature with height (at the end of the hour is displayed. For clarity, these profile curves consist of three types of lines, with one horizontal grid interval corresponding to 1 °C change of temperature. The topography is illustrated by a dotted line; the water sections of the Mediterranean Sea and the Dead Sea are indicated by heavy lines. Summer day.

cation is slightly unstable or neutral, $p \approx 0.1$ is typical for the 15 m level while it is generally somewhat lower at the 50 and 100 m level. The rougher surface on the coastal plain causes a slightly higher $p$ value there. The Mediterranean warm downslope flow,
penetrating over the lake (1800 LST), creates a stably stratified surface layer which is reflected by an increase in the computed \( p \) values.

The typical daily averaged \( p \) values for the areas of large daily averaged WP locations are typically in the range of 0.1–0.2, usually deviating only somewhat from \( p = \frac{1}{7} \) as suggested by Peterson and Hennessey (1978), although the simulated terrain and the associated flows are highly complicated. Over the water bodies (with their much smaller roughness), typical \( p \) values are around 0.05.

Many of the available surface wind data are from one low mast height. It is also assumed that low masts are typical in comprehensive WP observational surveys. For these reasons, we adopted a 5 m height as the reference level for the \( p \) computations. However, for the summer-day case, the possible effects on \( p \) due to the change of the reference height to 10 m, was tested. Since the difference, \( p_5 - p_{10} \), might have non-smooth variations with height, a detailed illustration, based on the model predictions with the high resolution vertical grid, is presented (Figure 6). It is evident that for the nocturnal case (0500 LST) over the inland domain, low \( p_5 - p_{10} \) values are associated with most of the unidirectional flow over the eastern mountain slopes, while the largest values are associated with the directionally sheared wind over the western mountain slopes, and along the coastal plain. The \( p_5 - p_{10} \) values decrease throughout the inland domain when unstable stratification is established during the daylight hours (1100 LST).

![Fig. 6. Vertical cross-section of the \((P_5-P_{10}) \cdot 100\) field (the differences in \( p \) values resulted by changing the reference height, \( Z_r \), from 5 to 10 m). Asterisks indicate the vertical model levels in the lower 100 m as adopted for this simulation. Summer case.](image)

6.2 Advective Sharav

During the spring and the fall, synoptic low pressure systems frequently prevail over the region resulting from the eastward migration of thermal synoptic lows from the Mediterranean coast of North Africa (Figure 3b), or due to a trough expanding northward from the Red Sea (Figure 3c), e.g., Levi (1967), Winstanley (1972). These situations are
characterized by warm and dry easterly flow in the lower atmosphere, causing a stable thermal stratification over the relatively cold water bodies. During daylight hours the high amount of incident solar radiation, resulting from the dry atmosphere, causes the development of a deep PBL.

Mahrer and Segal (1979b) have reported on the general resemblance of model-simulated advective Sharav fields and typical observed patterns over central Israel. We have adopted similar meteorological conditions for model initialization.

As shown in Figure 7, at the beginning of the night (2100 LST), when thermally induced winds are only slightly developed, flow patterns are affected primarily by the dynamic forcing of the topography (i.e., decelerated flow along the eastern slopes, and

![Diagram](image)

Fig. 7. Same as in Figure 4, except for advective-Sharav.
accelerated flow along the western sides and the coastal plain). Hence, this pattern reveals the dynamic effect of topography on the enhancement of WP over that available in the given synoptic flow. The development of nocturnal flows (0500 LST) increases the WP maximum west of the mountain ridge and tends to shift it offshore. The establishment of an onshore mesoscale sea-breeze component due to surface heating after sunrise, however, gradually eliminates the nocturnal peak. By 1300 LST, the establishment of a mesoscale thermally-induced pressure gradient counteracts the

Fig. 8. Same as in Figure 5, except for advective-Sharav.
synoptic flow and results in very light flows over the western part of the domain. Only at the mountain top does easterly flow still generate somewhat significant WP. Thus, for this type of synoptic flow, most of the substantial daily averaged contributions to WP occur at night and during the morning hours.

The effects of unidirectional and opposing synoptic and mesoscale flows on the vertical distributions of $p$ values at the termination of the nocturnal period is evident also in the present simulation (0500 LST). The values of $p$ are almost constant with height when the direction is nearly constant, while they change with height when it is not (Figure 8). A $p$ value of about 0.08 is typical in the highly mixed inland unstable surface layer (1300 LST), while the creation of strong thermally stable layers along the coastal area causes a significant increase of $p$ values. A sharp surface nocturnal inversion occurs on the second night because of large surface cooling associated with the very dry atmosphere (2300 LST). Although the thermal stratification is much more stable than computed at 0500 LST on the previous night (since no preceding history of Sharav conditions is considered in the prescribed initial data), the differences between the $p$ values at 50 and 100 m at these two times are generally small. Over the western inland domain, the nocturnal drainage flow is not completely established during the beginning of the second night, resulting in $p$ variations with height.

Since the daily averaged $p$ values are derived using daily averaged wind speed, they are weighted by the wind speed values. Therefore, because the nocturnal surface flow is much stronger than it is during daylight hours, the daily averaged $p$ patterns resemble the nocturnal type.

6.3 Winter Day

During the winter, the region is affected by mid-latitude pressure systems. Hence, light wind periods of several days associated with high pressure systems are followed by relatively intense surface winds, associated with low pressure systems travelling in the synoptic westerlies. Typically, prior to the passage of a low (Figure 3d), a strong surface southwesterly flow, veering clockwise with height, occurs (Atlas of Israel, 1963). Sometimes this flow pattern persists for periods longer than 24 h. This situation has been chosen to illustrate the winter day case.

Kessler and Pielke (1981) have evaluated the nature of inertial oscillations associated with strong flow perpendicular to adjacent narrow ridges. Such oscillations appear to play some role in the temporal variations of WP in the winter day case simulation (Figure 9). Generally, the maximum value of wind speed is found around 0300 LST (after 7 h of simulation which were preceded by 6 h of dynamical initialization), and a minimum at 1500 LST. The period of the flow oscillations, which is about 25 h, is of the same magnitude as that computed in the simple inertial case where advective terms, pressure gradient and turbulence are not considered, i.e.,

$$T = \frac{12}{\sin \phi}$$

(7)
where \( T \) is the oscillation period in hours and \( \phi \) is the latitude. In the present study, \( \phi \approx 32^\circ \), implying that \( T = 22.6 \) h.

The 0500 LST (about an hour before sunrise) WP is still affected by the peak of the inertial oscillation. However, the following mesoscale effects are evident:

(1) High WP is available over the Mediterranean offshore area due to the reduced friction surface.

(2) Along the coastal plain the effect of thermal induced nocturnal flow, along with the larger surface roughness, significantly reduces the WP.
(3) The maximum WP is located at the mountain top, while high WP occurs, and persists, along the upper windward slope of the mountain.

During daylight hours (1500 LST), the thermal difference between the land and the sea surfaces at this time of year is not sufficient to perturb the flow pattern significantly near the coast. Reestablishment of the nocturnal flow (2300 LST) is evident. However, the flow intensification toward the second maximum peak in the inertial oscillation influences the patterns.

Elbashan (1976) has evaluated surface wind speeds (at ~ 10 m height) for very strong winter storm situations over Israel. His evaluations are for synoptic conditions similar

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**Fig. 10.** Same as in Figure 5, except for winter day.
to those in the current winter day simulation, except that he considered stronger wind speeds. The upper curve in the daily averaged WP figure (Figure 9) illustrates the relative changes in wind speed along part of the simulated cross-section, as derived from Elbashan’s study. This curve displays a qualitative resemblance to the model-predicted variations of WP at 5 and 15 m levels.

The nocturnal p patterns (Figure 10) consist of high values at the coastal plain where stratification is stable because of the onshore advection of air masses which had been warmed over the sea. (However, if the advection of air is relatively cool, initially, and its transit over the water is short, thermal instability characteristic of conditions over the land with lower p values would be possible.) The p values reduce inland, where the stratification is less stable and surface roughness is reduced. In the slightly unstable stratification of the afternoon (1500 LST), the p values are slightly below 0.1, increasing again with the onset of stable stratification during the second night (2300 LST).

7. Discussion

The present study has demonstrated some of the potential use of numerical mesoscale simulations over inhomogenous terrain for both basic and applied WP studies. In terms of applied work, during periods of persistent steady synoptic circulation patterns, such as found over central Israel during the summer, the model can be a useful quantitative tool for the evaluation of the spatial distribution of WP. Even when the synoptic situations are much less persistent, such as over Israel in the fall through spring, the model can be a valuable qualitative tool in the assessment of wind energy resources.

This study has also demonstrated the substantial influence of mesoscale circulations on the temporal and spatial variability of wind power. Table I gives the WP values computed from the initial wind speed (see Figure 2). The values represent the expected WP if the initial synoptic wind alone was used as an estimate (namely, without considering the mesoscale forcing effect on the flow). The large deviation of our results from the values given in this table illustrates the importance of mesoscale perturbations on WP availability.

<table>
<thead>
<tr>
<th>TABLE I</th>
<th>WP values (W m⁻²) based on the initial synoptic wind speeds</th>
</tr>
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<tbody>
<tr>
<td></td>
<td>Simulated Case</td>
</tr>
<tr>
<td>Summer Day</td>
<td>1</td>
</tr>
<tr>
<td>Advective Sharav</td>
<td>36</td>
</tr>
<tr>
<td>Winter Day</td>
<td>107</td>
</tr>
</tbody>
</table>

The following two points refer to the advective Sharav and winter day simulations:
(1) Since WP ∼ V³, even a slight change in the wind speed can produce a relatively significant variation in WP when V is sufficiently large. For both of the aforementioned simulations, sensitivity tests have indicated that moderate modification in the prescription of initial wind speed (from that given in Figure 2), causes significant changes
in WP values. Qualitatively, however, WP spatial and temporal variations are affected only slightly.

(2) The lower atmospheric thermal structure of air masses advected into the region, as determined by the initial temperature profile, has some effect on the predicted wind energy characteristics. Therefore, an additional synoptic subclassification is needed in order to evaluate such effects.

Referring to p values, model predictions indicate several interesting features. During daylight hours, unstable stratification is predicted at inland locations most of the time, in all three simulations. These regions are characterized by inland p values which are almost spatially homogeneous, indicating the dominance of vertical turbulence on the vertical distribution of wind speed.

In the three simulated cases during the nocturnal flow regime, a decrease with height of the p values is typically predicted inland where mesoscale-induced flows cause a directional wind shear in relation to the prevailing synoptic flow. Over the lee slopes, where a stratified uni-directional flow is established, p, is generally, constant with height. Additionally, during nocturnal regimes over and near mountainous terrain, stable stratification has been found to be associated with relatively low p values (compared with those commonly specified for homogenous terrain cases). This indicates that mesoscale pressure gradient and advective effects are dominant influences affecting the vertical profile of the horizontal wind.

Finally, daily averaged p values at the locations of maximum averaged daily WP are about \( \frac{1}{3} \), suggesting the use of this value to estimate WP over mountainous terrain when observational estimates of p are unavailable.

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