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Determination of actual evapotranspiration and transpiration in desert sand dunes (Negev Desert) using different approaches

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Abstract In an arid environment, especially in sandy areas where surface runoff is of no practical importance in the hydrological budget, it is rainfall, dewfall and evapotranspiration that become the most important variables. To assess actual evapotranspiration, several methods (flux-gradient, BREB, eddy correlation) were applied to data from the Nizzana experimental site in the northwestern Negev Desert. Additionally, a model specifically designed for arid environments is introduced in this paper. This zero plane model shows the most reasonable results compared with the other methods, which overestimate evapotranspiration to a large degree. It is shown that plant transpiration is the dominant process in total evapotranspiration while advective processes do not play a major role in the near-ground boundary layer, although the study area is influenced by a sea breeze. Actual transpiration of *Artemisia monosperma* was measured in a field experiment to validate the calculated evapotranspiration. The vegetation contributed 41% of the calculated total evapotranspiration in a single month.

Key words desert microclimate, arid environment, desert, sand dunes, *Artemisia monosperma*, Nizzana, Negev

1 Introduction

The knowledge of atmosphere, plant and soil interaction in the sand dunes is limited. There is a need for a better estimation of water flows and evapotranspiration in dry lands. A detailed water budget model is important to optimise shelterbelt design in arid and semi-arid regions in the future (Veste and Littmann, 2005). The potential vegetation cover is determined by the water resources available for the plants and their water use strategies (Li and Shi, 2003; Littmann and Veste, 2005; Veste et al., 2006).

Several evapotranspiration models have been developed and tested in various ecosystems (e.g. Herbst et al., 1996; Malek et al., 1997; Domingo et al., 1999, 2001). The most widely used model is the single-source Penman-Monteith evapotranspiration model (Monteith, 1965) which assumes that canopies can be regarded as one uniform surface or big leaf. In the case of the relative homogenous and not drought-stressed agricultural fields the big leaf model results in a good estimation for evapotranspiration and scaling from leaf-level gas exchange to canopy scale can be successful (Baldocchi, 1993). Ventura et al. (1999) found in agricultural fields that different evaporation equations provide in general acceptable results, but the study demonstrated that under special climatic conditions the Penman-type model sometimes failed. In more complex topographic regions and vegetation stands (e.g. forests, mountains) the various microclimatic approaches will lead to false estimation of the evapotranspiration. In contrast to humid ecosystems the arid and semi-arid regions are characterized by patchy vegetation with larger open spaces. Water vapour pressure fluxes are generally low when compared to the humid environment and saturation deficits are high, both of which may lead to an estimate of potential, not actual evapotranspiration in the Penman approach.

In this paper we determined the evapotranspiration in a sand dune ecosystem in the northwestern Negev desert using different microclimatic approaches. A modified approach was developed for the estimation of evapotranspiration in arid and semi-arid sparsely vegetated areas. For a comparison with the calculated results we measured independently from the microclimatic approaches the transpirational water loss of the vegetation.

2 Material and methods

2.1 Study sites

The investigations were carried out in the sand dunes of the northwestern Negev (Fig. 1). These sand dunes are the eastern most of the sand field covering the northern part of the Sinai Peninsula and the north-

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western Negev (Veste et al., 2005). The local climate is determined by a sharp gradient from the Mediterranean coast to the arid climate of the Negev. The average annual rainfall decreases from around 170 mm at the northern edge of the sand field near Yevul to approximately 90 mm near Nizzana. Rainfall is limited to the winter season from October to March. At the Nizzana experimental site the rainfall record since 1991 indicates a period of 3 years between two above normal rainy seasons, i.e. > 90 mm in 1991/92 (131 mm) and in 1994/95 (148 mm) while the period in-between was average or below average (87 mm in 1992/93 and 50 mm in 1993/94). Following the very good rainy season of 1994/95, in 1995/96 rainfall was very low (38 mm) and was only average in 1997/98 (78 mm). Predominant shrub species of the sand dunes are *Anabasis articulata*, *Artemisia monosperma*, *Thymelaea hirsuta*, *Convolvulus lanatus*, *Moltkiopsis ciliata*, *Echiochilon fruticosum* and *Retama raetam*. The mean vegetation cover in the interdune area is around 20%, on the dune crest 16% (Littmann and Veste, 2005). The northern site near Yevul is dominated by *Artemisia monosperma*. The vegetation cover is 22% and varied between 12% and 35% (Veste et al., 2005).

Fig. 1 Location map of the investigation sites Nizzana and Yevul in the sand dunes of the northwestern Negev (Israel)

2.2 Microclimatic measurements

Microclimatic measurements were carried out at two study sites in the northwestern Negev at the southern margin near Nizzana and the northern margin near Yevul. In the most southern site in Nizzana two recording stations were installed on the south and north facing slopes and data for the model development were recorded from July 1995 to June 1996. The northern station is located in an *Artemisia monosperma* semi-shrubland. Measured variables (hourly means) relevant for this investigation include net radiation at 2 m above ground (REBS Q7.1-L net radiometer, Campbell Scientific Ltd., Logan, Utah, USA), soil heat fluxes in –0.15 m below ground (heat flux plate, REBS HFT3-L, Campbell Scientific Ltd., Logan, Utah, USA), air temperatures at 2 and 0.2 m above ground (Pt 100 thermoresistor probes, Campbell 107-L, Campbell Scientific, Logan, Utah, USA), relative humidity at 2 m height (Vaisala HMP35C-L, Helsinki, Finland) and horizontal wind speed at two heights (0.15 and 2 m, mini cup anemometer 03101-L, R.M. Young, Traverse City, Michigan, USA). Measurements were recorded every 5 s by a data logger (CRX 10, Campbell Scientific Ltd., Logan, Utah, USA) and averaged every hour. In this paper we present only the data for the north facing slope. Vertical wind speed was measured only on the south facing slope; therefore the eddy correlation technique was only applied to this slope. This conventional arrangement of field measurements, especially the time resolution of averaging intervals, necessarily leads to some inaccuracy that affect the model results with regard to highly time-dependent approaches like the eddy correlation method.

2.3 Evapotranspiration models

Most approaches to approximate actual evapotranspiration are either vapour flux models, energy balance models or physical-empirical models (Schrödter, 1985). A vapour flux model may be based on gradients of properties of a near-ground air column following Fick's law of diffusion, such as wind speed and specific humidity. Introducing an explicit form of the turbulent diffusion coefficient of momentum, Thornthwaite and Holzman's (1942) gradient-flux model considers evapotranspiration as the upward (positive) flux of specific humidity between two heights above ground enforced by vertical wind shear:

$$ETP = \rho k \frac{(q_{z1} - q_{z2})(u_{z2} - u_{z1})}{\ln\left(\frac{z_2}{z_1}\right)^2} \qquad (1)$$

where u_{z1} and u_{z2} are the horizontal wind speeds at heights z_1 and z_2; q_{z1} and q_{z2} are the respective specific humidities; ρ is the air density and k the Karman constant. Since this method requires a determination of the logarithmic wind profile above the roughness height z_0 which in turn is the aerodynamic surface where turbulent vapour flux may start, we computed mean monthly z_0 from best-fit hourly adiabatic and diabatic wind profiles between 2 and 0.5 m including the Richardson number as stability parameter. The determination of neutral, stable or labile situations followed near-ground temperature profiles. In our case z_0 remained constant at 0.19 m at the north facing dune slope, which coincides with the lower height of conventional temperature measurements. The actual wind profile for each hour was then computed from 2 m down to z_0. Since no pressure measurements were available, humid air density was computed using standard pressure. However, we found that even large pressure fluctuations will not have large effects on the respective values. The same applies to the calculation of specific humidity from relative humidity and saturation pressure. For the 0.2 m level, air temperature at this height was used for the respective computations.

A direct method to estimate evapotranspiration is the eddy-correlation technique (Swinbank, 1955). Based on the fact that turbulent exchange of properties in the boundary layer is triggered by the friction velocity, surface roughness and gustiness in terms of Reynold's covariance, which considers the actual wind speed as the sum of mean wind speed over a period of time and the momentary variance, the approach correlates the vertical fluctuations of wind speed and specific humidity over a certain period:

$$ETP = \overline{(\rho\omega)' q'} \tag{2}$$

where $(\bar{\rho}\,\bar{\omega})$ is the mean vertical flux of mass, $(\overline{\rho\omega})'$ and $\bar{q}'$ are the respective mean deviations of mass flux and specific humidity from their means over the observation period.

Including the energy budget and a parameterisation of turbulent flux, the Bowen Ratio Energy Balance (BREB) technique originally presented by Sverdrup (1936) is widely accepted as the physically most sophisticated approach (cf. Schrödter, 1985):

$$L_v E = -\frac{R_n + G}{1 + \gamma \dfrac{\partial t / \partial z}{\partial e / \partial z}} \tag{3}$$

where R_n is the net radiation, G the soil heat flux, γ the psychrometric constant (0.66 hPa·K^{-1}) and e the water vapour pressure (hPa). $\partial t/\partial z$ is the temperature difference between two heights of measurement and $\partial e/\partial z$ the respective vapour pressure difference. L_v is the latent heat of vaporization (2.45×10^6 J·kg^{-1}) and E the evapotranspiration. Eq. (3) divided by L_v converts the flux of latent heat into actual evapotranspiration values (mm).

For the purpose of assessing the plausibility of the different model outputs it is feasible to use the potential evapotranspiration as an upper limit value. In the context of potential evaporation from open water surfaces the combined method after Penman (1948) has been applied to a wide range of world climates. Strictly physical, the approach includes a parameterisation of the energy balance and the influence of wind:

$$ETP_{pot} = \frac{s}{s+\gamma} R_n + \left(1 - \frac{s}{s+\gamma}\right)(E-e)\, 0.27 \left(1 + \frac{ul}{100}\right) \tag{4}$$

where E is the saturation vapour pressure, s vapour pressure and ul the mean daily wind length. Also developed for non-arid climates, Penman's rather simplified wind function will lead to a larger distortion of results in case of high surface roughness or advective processes that obliterate the diurnal course of wind speed and water vapour in the boundary layer (Schrödter, 1985). Using a daily mean sum of positive net radiation, the Penman approach applies only to daily means of evapotranspiration and is not suitable for a higher time resolution in its present form. There have been efforts to develop correction factors for various climates (Doorenbos and Pruitt, 1984) or to define monthly minimum thresholds for humid areas (Penman, 1956). However, the application of the combined Penman formula to irrigated land in an arid region (Ohlmeyer and von Hoyningen-Huene, 1975) showed that the results for potential evapotranspiration even underestimated actual values by 27%.

2.4 The zero plane model

One of the main problems of approaches using vapour gradients between two heights of measurement originates from the fact that vapour fluctuations resulting from evaporation or condensation processes will be much larger just above the active surface as compared to a greater height. Using the mean specific humidity of the air column under consideration or its gradient between two heights, without including air movement as a purely empirical boundary condition, will underestimate actual evapotranspiration and especially dewfall. The inclusion of wind speed or wind shear as in Thornthwaite-Holzman will lead to overestimations. On the other hand, gradient-free approaches like the eddy-correlation technique will fail in relieved terrain when turbulent vapour flux is enforced by topographic obstacles.

Alternatively, the passage of vapour through a two-dimensional plane near the active surface, which may be the canopy layer in case of vegetation stands, the bare surface, or, most likely, a height in-between both reflects, depending on its direction, either evapotranspiration or potential condensation in terms of a fluctuation of properties (i.e. sensible and latent heat) as a function of time. We assume that all processes (stability, turbulence, advection, evapotranspiration, dewfall) will effect fluctuations of specific humidity on such a near-surface plane at the relative bottom of the air column involved in the processes, irrespective of the actual height or volume of the column. However, any such approach requires vertical fluxes that are relatively free of advective disturbance. Advection may be excluded only in case of a measurement area large enough to minimise divergence of the humidity field as in most humid regions (Schrödter, 1985). Especially in arid regions the influence of advection cannot be neglected easily as it may lead to either an increase in actual evapotranspiration as compared to the water equivalent of the radiative budget because of a much larger saturation deficit (cf. Ohlmeyer and Hoyningen-Huene, 1975) or a decrease when moist air is advected, e.g. in case of sea breezes as present in our study area during the half year of summer (Littmann, 1997). In an arid region of China, Niu et al. (1997) found, by means of a coupled soil-atmosphere model, the vertical flux of latent heat to control specific humidity up to a height of 100 m.

Naturally, vapour or dry air advection is coupled to fluctuations in horizontal and vertical wind speed on different time scales. Paw et al. (1995) reported that the advective term in energy exchange within a canopy layer is represented by high frequency scalar traces such as temperature change within the 10^0 to 10^{-1} s time range. However, advective effects will be largest in case of frontal passages or khamsinic depressions over the area. To exclude part of such disturbance we did not consider cases that showed anomalous wind speeds relative to the mean monthly diurnal course of wind speed (roughly 10% of the overall series) following the anomaly filter:

$$\left(u_{Tn} - \bar{u}_{Tn}\right) > \sigma_n \tag{5}$$

where u_{Tn} is the actual data n at time T of the series, $\bar{u}_{Tn}$ is the mean at time T and σ_n is the standard deviation.

As fluctuations occur in terms of variation around a mean value, we chose to split the monthly mean of specific humidity at the measurement height of 0.2 m into 24 hourly means in order to observe vapour fluctuations in terms of explained variation in-between these group means. The following model was developed in three consecutive steps (Fig. 2):

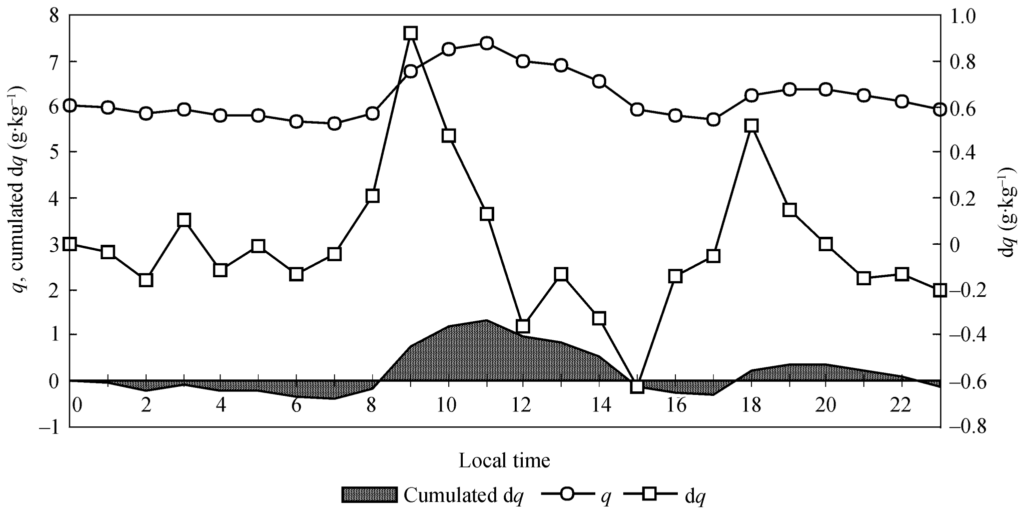

Fig. 2 Evolution of the zero plane model. Mean hourly values of specific humidity (q) over a month in the plane of measurement are differentiated (dq) and dq is expressed as a cumulative series with a zero start value. The shaded area is used for integration of mean monthly specific humidity fluctuations in the zero plane.

1) Vapour fluctuations within the plane of observation (strictly controlled by evapotranspiration and condensation processes) are expressed as a differential series of the mean hourly values of specific humidity dq. They may be positive in case of an upward vapour flux from the active surface of evapotranspiration (soil and canopy) through the plane while an opposite flux from above will not be detectable as positive dq because it may occur in condensation situations only (except in rare cases of strong turbulent advective downward mixing). Decreasing dq either reflects the condensational removal of water vapour from the near ground air column or constantly decreasing vapour input from the active surface. In fact, we found that dq at 0.2 and 2 m are closely interrelated (r^2 = 0.78) with no time lag but at 0.2 m, i.e. close to the active surface, dq is much more variable than at 2 m (σ = 0.9 vs. 0.7). The use of hourly means excludes high-frequency turbulent signals that would result in no vapour fluctuation on the 10^0 to 10^{-1} s time scale in many cases because of equalling upward and downward passages (Jacobs, pers. comm.).
2) Actual evapotranspiration should be the total amount of water vapour passing through the plane at time intervals ΔT in terms of an increase in specific humidity. In the case of dewfall on the

other hand it measures the total amount removed from the plane. However, removal or increase within the mean time series sets in relative to the prior fluctuation. This is represented by the cumulated dq series and is consistent with the principle of continuity since cumulated dq also represents the flux of latent heat through the plane. A similar approach including the introduction of finite differences and summation of such derivatives with reference to heat exchange in canopies was introduced by Paw et al. (1995).

3) Finally, the cumulated series is integrated following the determination of precipitable water (Peixoto, 1973):

$$ETP = z\rho_w \int_{T=1}^{T=24} f\left|(\bar{q}_{(1..24)} - \bar{q}_1)\right| > 0 \qquad (6)$$

where z is the unit height of the plane (1 m) and ρ_w is the density of water vapour. We use the functions of the differential specific humidity series (which show a start value of zero at 0:00 local time, hence the "zero plane") because it is identical with the cumulated dq series. Eq. (6) includes the value of the integrated functions since it may happen that the differential series leads to negative values during daytime when the specific humidity decreases relative to the zero start value. This is the case when either dry air is advected during khamsinic situations in spring (March, April, May), not detectable by the anomaly filter (5) or transpiration is greatly reduced in the afternoon (summer). Nighttime negative values following (6) are subject to the critical dewpoint filter and we found a perfect coincidence of hours where condensation is physically possible (i.e. where the mean frequency of cases with dt_d < 3 K is high until it drops to zero during the first hour after sunrise when net radiation values become positive) and those where the model output indicated a decrease in specific humidity.

2.5 Transpiration measurements

For this experiment a stand predominated by *Artemisia monopsperma* was chosen to allow a better estimation of the area-related transpiration calculated from the leaf gas exchange data. The measurements were conducted in the northern experimental site near Yevul (Fig. 1). Diurnal courses of transpiration were measured at 8 days in March 1999 using a portable porometer system (HCM-1000, Walz GmbH, Effeltrich, Germany). The water vapour exchange was analysed with an infrared gas analyser (BINOS 100, Rosemount, Hanau, Germany) and water fluxes were calculated using the DA-1000 software (Walz, 1996). The porometer head is climatized with Peltier-elements and the climatic conditions inside the cuvette corresponded with the ambient air temperature and water vapour deficit (vpd). This is important in order to avoid overheating of the cuvette resulting in overestimation of transpiration (Midgley et al., 1997). For the extrapolation of the leaf area based transpiration rates to soil area based fluxes all leaves of *Artemisia monosperma* in a representative plot of 5 m×5 m were harvested and the leaf area index was calculated with LAI = 0.12 m·m^{-2}. Precipitation at the experimental site during the experiment was 13 mm rainfall and 2 mm dewfall.

3 Results

3.1 Evapotranspiration models

The monthly mean evapotranspiration from the Nizzana experimental site are shown in Fig. 3.

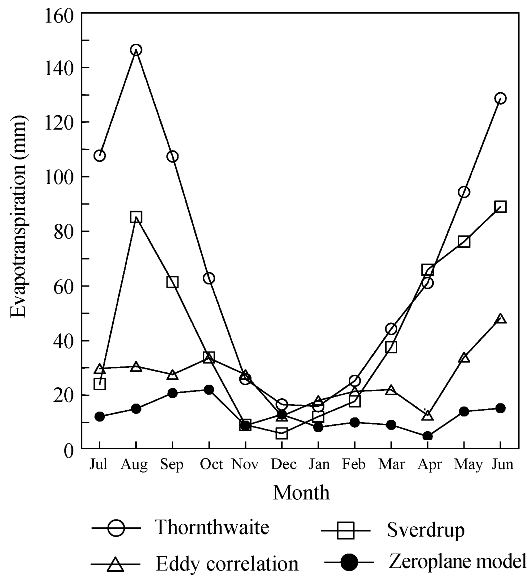

Fig. 3 Monthly evapotranspiration in Nizzana (1995/1996) using different model approaches

All approaches show fairly similar values for the winter months (November to February). Over the summer months, however, deviations became extreme. It is especially the Thornthwaite-Holzman method which models monthly totals that are close to potential evapotranspiration after Penman. Based primarily on the energy budget, the Sverdrup (BREB) model shows a very similar annual course with an extremely low value in December but also fairly high summer totals because it will yield higher water equivalents when both net radiation and vapour gradients are high as in summer. Except for winter the second model group does not show much similarity to those two methods. Both the eddy-correlation technique model and the zero plane model are not dependent on the height of the air column considered in the gradient approaches and both methods use mean values over the observational period. Three values (July to September) were filled in by regression techniques due to lack of vertical wind speed data. Although the assessment of such data was quite limited because of the inadequate time resolution of the wind speed data, except for December when the eddy correlation model constantly showed higher results than the zero plane model, it equally included vertical wind speed fluctuations in terms of turbulent transport, while the zero plane approach considered only the actual specific humidity fluctuations. Thus, there is a wide range in the model results for annual evapotranspiration (Fig. 4).

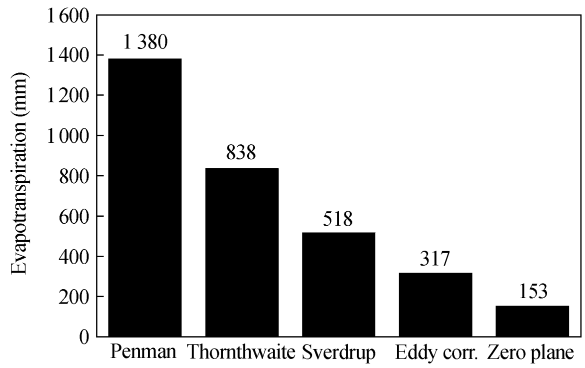

Fig. 4 Annual evapotranspiration at the Nizzana site for the period July 1995–June 1996 using different model approaches

3.2 Zero plane model

The zero plane model results for January 1996 and July 1995 are shown in Figs. 5 and 6. Specific humidity decreases from the zero start value until sunrise and the onset of labilization during the first hour thereafter, which is also matched by the dewpoint filter frequencies and by the frequency of near-ground stability. In both examples, as well as in all individual months, specific humidity shows a sharp increase for the first three hours after sunrise towards the daily maximum at 8:00 to 10:00 or 10:00 to 12:00, depending on the season. This maximum is well before the saturation deficit maximum which indicates that actual evapotranspiration is not simply controlled by the saturation deficit (analysis of variance results in 13% overall explanation of specific humidity fluctuations in terms of the saturation deficit). However, in the afternoon when the saturation deficit is highest and also the near-ground wind speed reaches its maximum water vapour decreases to values near or well below the zero start value except for a small secondary peak in July. After sunset specific humidity increases again to a level well above the start value parallel to the development of a stable layer. Wind speeds at 2 m show a close interrelation with wind directions (66% explained variation in one-way analysis of variance), which explains the diurnal course of wind speed in terms of land and sea breezes, while specific humidity does not (5% explained variation by both wind direction and wind speed). This implies that periodic daily advection by the regional wind systems does not play a major role.

3.3 Transpiration

Diurnal courses of the microclimatic and transpiration measurements are shown in Fig. 7. The measured mean daily water loss of *Artemisia monosperma* calculated from the gas exchange data was 1.6 $kg \cdot m^{-2}$ leaf area per day. Therefore, the estimated water loss of this semi-shrub stand was 5.8 mm per month in March 1999 and the modelled monthly evapotranspiration was 14 mm. The transpirational water loss of the *Artemisia*-shrub community contributed 41% of the actual evpapotranspiration calculated by the zero plane model.

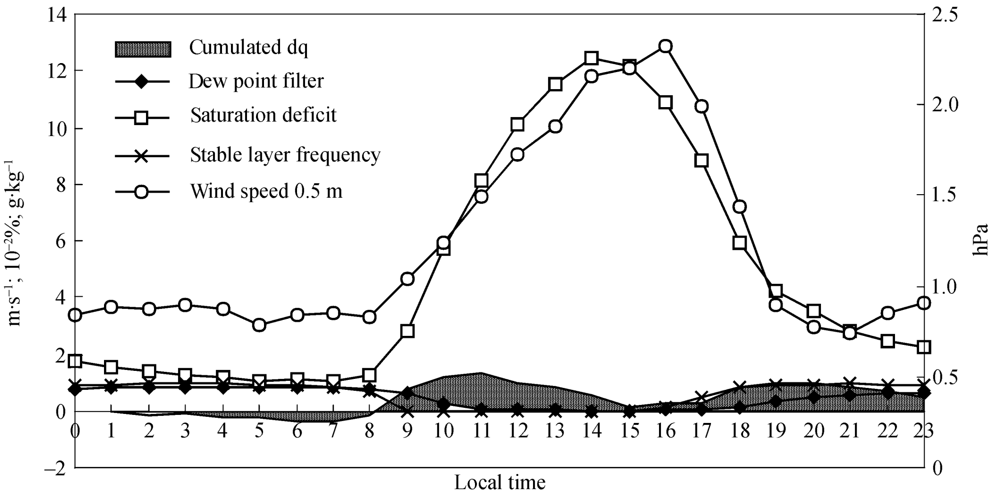

Fig. 5 Zero plane model output for January 1996. Note that cumulated dq ($g \cdot kg^{-1}$), dewpoint filter and stable layer frequencies (in 10^{-2}%) and wind speed on 0.5 m ($m \cdot s^{-1}$) are on the left y-axis; saturation deficit (hPa) is on the right y-axis.

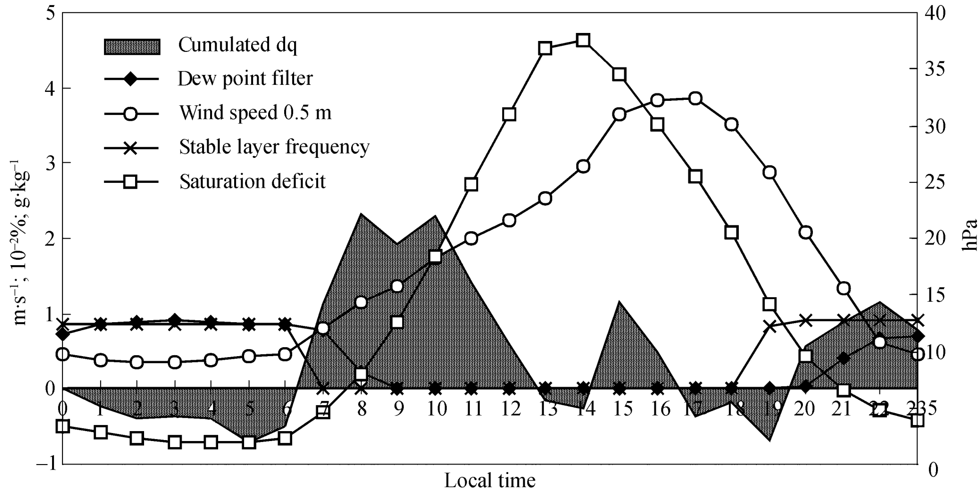

Fig. 6 Zero plane model output for July 1996. Note that cumulated dq (g·kg^{-1}), dewpoint filter and stable layer frequencies (in 10^{-2}%) and wind speed on 0.5 m (m·s^{-1}) are on the left y-axis; saturation deficit (hPa) is on the right y-axis.

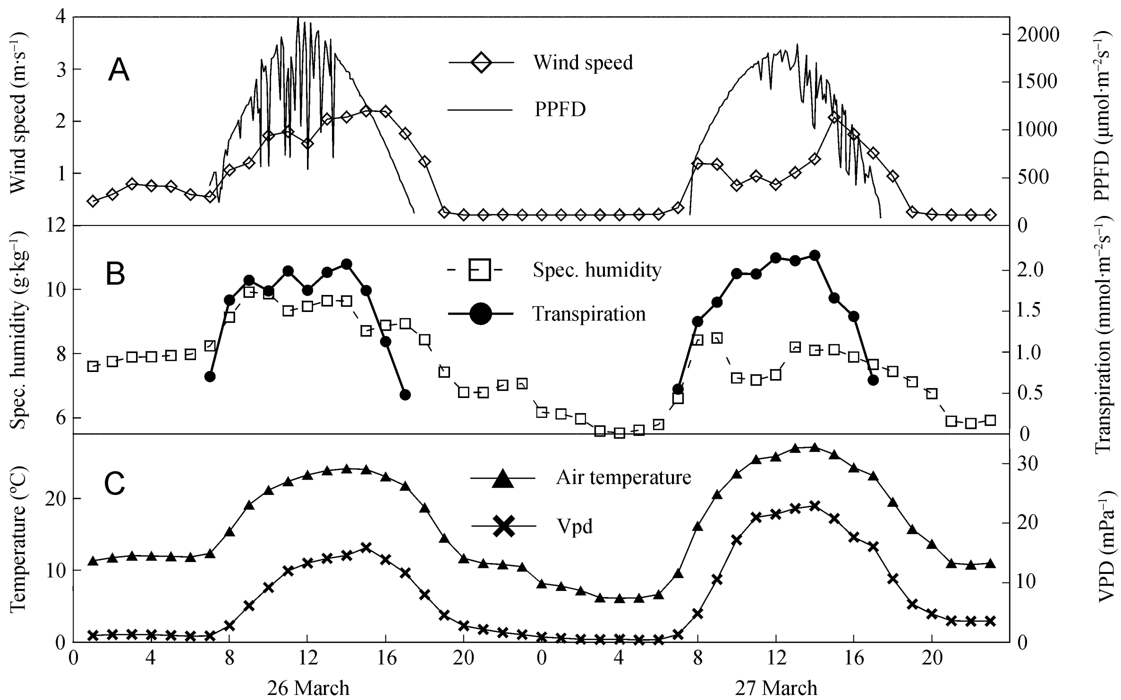

Fig. 7 Daily courses of microclimate parameters (A: wind speed, PPFD; B: specific humidity; C: air temperature and vpd) from the microclimate station and transpiration (B) of *Artemisia monosperma* measured with a porometer system at Yevul (March 26–27, 1999)

4 Discussion

The various explicit models applied to the study of actual evapotranspiration in an arid environment provide a wide range of results. We showed that evapotranspiration is largely controlled by plant transpiration while surface evaporation occurs only after rainfall or, secondarily after dewfall, in the study area.

Since wind speed, wind direction and the saturation deficit all explain but 20% of total specific humidity variation in the zero plane model it is apparent that the daytime fluctuations are mainly controlled by the diurnal cycle of transpiration typical for most desert plants. After sunrise transpiration starts, but may be actively reduced by closing the stomata when the air to leaf vapour pressure difference increases. During and after a good rainy season there was no drastic stomata midday depression for various shrubs in the Nizzana sand dunes (Veste and Breckle, 1996a, 1996b). In this cases, water uptake from the lower root zone of the plants enabled higher transpiration rates. After low winter rainfall and during the dry season, however, the shrubs showed stomata closure as well as a gradual reopening in the late afternoon (Veste and Breckle, 1996b). In this way, measured diurnal courses of plant transpiration match water vapour fluctuations on the zero plane to a high degree. However, water vapour decreases on the zero plane in terms of a downward flux towards the surface of condensation only after the near-ground air temperature is low enough to drop the dewpoint temperature difference below the critical value, i.e. increasingly during the early morning hours before sunrise.

Further valuable results for the evaluation of our model results are provided in the study of rainwater percolation in the Nizzana dunes by Yair et al. (1997). They found that on a north facing dune slope in a year following an average rainy season, as in 1993, water percolation is limited to a depth of approximately 0.6 m while after a very good rainy season (1992) percolation reached down to 4 m with lateral subsurface flows on the north facing slope next to our recording station. Water content of the upper layer was around 3% in the drier year, in the lower layer 1% while it increased in the wetter year to 6% and 2% respectively. These data imply water storage of 18 mm in the upper root zone (–0.6 m) and of 35 mm in the lower one in a dry year but of 106 mm in the total rooted soil column in a wet year. Recharge of soil water in the lower root zone occurs only after a good rainy season (Yair et al., 1997). If we calculate from these data the time the accumulated soil water within the total root depth of the dominant plants (at least 3.5 m; cf. Vested and Freckle, 1996a) over three years in-between two good rainy seasons would be stored and apply the different model outputs (evapotranspiration in Table 1 minus approximately 30 mm of dewfall equalling surface evaporation), the water resources would have been used completely by the vegetation surrounding the recording station (*Artemisia monosperma* shrubs) after 3 months (Thornthwaite-Holzman), 5 months (Sverdrup), 8.6 months (eddy-correlation) and 20 months (zero plane).

However, since this did not occur, it allowed us to estimate the necessary root area over the rooted soil column and the respective vegetation densities that would be in equilibrium with the rainfall regime over the three year period under consideration. This is because root competition of the plants will lead to a projected root area that allows water uptake of a plant sufficient to endure the rechargeless period between two good rainy seasons while non-competitive individuals will die. The Thornthwaite-Holzman model would imply a necessary projected root area of 12 m^2 equalling a vegetation density of 8%, the Sverdrup model 7.2 m^2 (14%), the eddy correlation technique 4.2 m^2 (24%) and the zero plane model 1.8 m^2 (55%). In fact, the observed vegetation density on the north facing slope at the Nizzana site ranges from 21% to 30% and can be in spots at the upper slope up to 40% (Littmann and Veste, 2005; Veste et al., 2005).

The measured transpirational water loss of the *Artemisia*-shrub community with 41% showed also realistic values and supports the estimation by the zero plane and eddy correlation models. Comparable data on transpiration in desert ecosystems are rare. The mean annual transpirational water loss of a plant community dominated by *Zygophyllum dumosum* on loess soils in the northern Negev south of Beer Sheva was around 75 mm per year (Zohary and Orshan, 1954), while the annual rainfall of the nearest meteorological station of Revivim was only 58 mm during the investigation period and 113 mm in the previous year (data Meterological Service of Israel). In the North American arid and semi-arid shrub community the percentage of transpiration attributable to total evapotranspiration varies from 34% to 54% (Caldwell et al., 1977; Smith et al., 1995; Reynolds et al., 2000).

5 Conclusions

We can conclude from the results that in an arid natural ecosystem the zero plane model should provide the most plausible results when compared to all other approaches. The eddy correlations values were within a plausible range, but their applications need a high technical input. Therefore, the zero plane model provides an alternative way for a correct estimation of evapotranspiration with standard microclimate equipment. But it should be clearly stated that the zero plane model would work only in dry lands where the diurnal and hourly differences of specific humidity, close to the active surface, are large especially when controlled by transpiration. In humid climates such a diurnal course is regularly obliterated by strong advective mixing and the model does not provide reasonable results. This was found when applying the model to comparable data from measurements in northern Germany where dq may be zero over long periods. On the other hand, the model is rather conservative as advective influences on the near-ground plane are quite limited and are largely removed from the series by the anomaly filter as compared to greater heights or other approaches using gradients.

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(BEO 0339495A, MV).

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