Diurnal Water Content Changes in the Bare Soil of a Coastal Desert

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ABSTRACT

The deposition of dew is a common meteorological phenomenon that has been recognized as an important ecosystem element, especially in arid areas. There is some evidence that indicates that there is an increase in the water content of the topsoil during nights in which no dew deposition was observed. The purpose of this study is to describe the daily pattern of changes in water content in the upper soil layers and to identify the mechanism by which water is added to the soil (deposition or direct absorption). Moreover, the gains in soil water content during the night are compared to the dew amounts recorded by the Hiltner balance, and the losses and gains of water in terms of easily measurable environmental parameters are parameterized.

Nine 24-h field campaigns took place during the dry season of 2002. During each campaign, the 100-mm topsoil was sampled hourly, and water content at 10-mm increments was obtained. Micrometeorological measurements included incoming and reflected shortwave radiation; net radiation; wind speed at four levels; dry- and wet-bulb temperatures at 1-m height; and soil heat flux. In addition, the changes in mass of an improved microlysimeter were recorded, and dew deposition amounts were measured using a conventional Hiltner dew balance.

The results of this study indicate that in the area in which this study was carried out actual dew deposition *on a bare soil surface* is probably a rare occurrence. There is, however, a clear discernible daily cycle of water content in the upper soil layers. The lack of any evidence of soil surface wetting led to the conclusion that the main process responsible for the observed diurnal change in water content is the direct adsorption of water vapor by the soil. A strong and significant correlation was found between the total adsorption of water vapor by the soil during the period that begins in the early afternoon and ends at sunrise and the total potential evaporation between sunrise and sunset of the previous day. Based on this finding an empirical model is proposed in order to predict the total amount of water adsorbed by the soil during the absorption period. The proposed model is probably site specific but is very simple and easy to implement. An additional outcome of the present study is that, in the area in which it was carried out, artificial condensing plates are poorly correlated to water vapor absorption, and the deposition of dew on them is not indicative of dew deposition on bare soil.

1. Introduction

The deposition of dew is a common meteorological phenomenon. The magnitude of the fluxes involved is very small, which would indicate a priori that the potential contribution of dew to the water balance of a given region would be very minor. Nevertheless, its contribution to the water budget of plants has been studied, and particular attention has been paid to the microclimate that develops within the plant canopy (Pedro and Gillespie 1982; Zuberer and Kenerley 1993; Wilson et al. 1999; Jacobs et al. 2000). It has been reported that in arid areas, dew is an important ecosystem input because of the lack of precipitation (Evenari 1985). It is a major source of water for the growth and development of biological soil crusts (Lange et al. 1992, 1998; Jacobs et al. 1999); it plays a key role in the germination of annual seeds (Gutterman and Shem-Tov 1997); it is an important source of water for small insects (Moffett 1985); and finally, it plays a role in the water balance of some vascular plants (Willis 1985; Zentay et al. 1985; Jacobs et al. 2000).

Dew is not only important in a biological context. Its presence, for instance, may affect remotely sensed parameters. It has been shown that altimeter backscatter measurements are affected by the presence of dew on the surface (Ridley et al. 1996) and that surface reflec-

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tance values were affected by the daily cycle of dew deposition and subsequent evaporation (Menenti et al. 1989). For these and similar applications it is enough to know the time span during which dew is present on the ground. Dew may, in addition, play an important role in deserts by changing the water and energy surface balances. It is a commonly accepted fact that the magnitude of latent heat flux above desert areas is linked to the occurrence of precipitation events and is negligible during the dry season during which the radiant energy reaching the surface of the desert is partitioned between convective and conductive sensible heat only (Cleugh and Roberts 1994; Unland et al. 1996). This pattern could be altered by the presence of a wet soil surface, which could happen if enough dew would condensate.

In order to assess the degree to which dew may affect the above-mentioned balances, it is necessary to quantify the actual amounts of water deposited and their distribution within the soil matrix. Because plants cover only a very small fraction of the surface of deserts, the study of dew deposition on bare soil surfaces is essential. This task, however, is challenging and introduces many difficulties, since the fluxes and the total amounts of water deposited are extremely small. To overcome these difficulties the usual approach has been to assess the condensation on detached surfaces (on which measurements are easier to perform) and assume that they are representative of the actual deposition on the substrate above which they are installed. A number of devices to determine dew deposition amounts have been proposed and are in use (Duvdevani 1947; Lomas 1965; Noffsinger 1965; Bunnenberg and Kuhn 1980; Zangvil and Druian 1980; Severini et al. 1984; Janssen et al. 1991; Jacobs et al. 1994; Zangvil 1996; Kidron 1998; Liu and Foken 2001), their common denominator being that the physical properties of the artificial condensing plates are very different from those of the bare soil.

Among these instruments, the Hiltner dew balance showed great promise because of its simplicity and robustness. It was used for several years in the Negev Desert in Israel in order to monitor dew deposition (Zangvil and Druian 1980; Zangvil 1996). This instrument consists of an artificial condensation plate (hanging from a beam 20 mm above the soil surface) that is continuously weighed. The energy balance of the condensation plate is completely different from that of the soil above which it hangs because of the following: 1) the plate is isolated from the soil surface by an air gap; 2) the properties of the material from which the condensation plate is made (a thin plastic plate) are very different from those of the soil; and 3) the dew condensing on the plate accumulates on it and may evaporate, while dew formed on the soil surface may be absorbed by the porous matrix. In view of these limitations (which also hold true for the other dew-recording methods or devices mentioned above) the Hiltner dew balance could be considered as a ''potential dew'' gauge whose results are mainly correlated to atmospheric conditions (Ninari and Berliner 2002).

The amounts of dew deposition, recorded in the Negev Highlands using the Hiltner balance, were in the range of 0.06–0.1 mm per night on average, and 200 dewy nights per year were detected (Zangvil 1996). Close by, in a very similar area (soil type, topographical elevation, etc.), Kidron (2000) estimated dew using a different type of gauge (Kidron 1998). This gauge was found to be well correlated with the Duvdevani dew gauge, but suffers essentially from the same set of drawbacks as described previously for the Hiltner balance. Kidron (2000) made the interesting comment that even though dew deposition was recorded by their dew gauges, no visible moistening of the soil surface was evident. This observation raises the interesting possibility that the detection of dew by gauges located on or close to the soil surface may not be indicative of actual dew deposition on the soil surface. This does not, however, preclude the possibility that water vapor from the atmosphere may be directly absorbed by the soil matrix as a result of capillary condensation and/or physical adsorption. The former is the predominant mechanism when the relative humidity in the pores is high, while the latter predominates at low values of relative humidity (Philip and de Vries 1957). Water adsorption has been put forward as being an important link in the water cycle of arid and semiarid regions (Danalatos et al. 1995; Kosmas et al. 1998, 2001).

Kosmas et al. (1988) proposed an empirical model to predict the water vapor adsorption as a function of the minimum daily relative humidity, the daily amplitude of relative humidity, and the soil water tension of the 50-mm topsoil. The usefulness of such a model is limited, however, as water tension of the uppermost soil layer is rarely measured, and most definitely not on a regular basis.

The previously mentioned studies (Zangvil 1996; Kosmas et al. 1998; Kidron 2000) indicate that there are a large number of nights during the dry season during which, irrespective of the mechanism (dew deposition or direct vapor absorption), water is added to the bare soil surface in arid environments. The added water will evaporate the following day, thereby changing the pattern of radiant energy dissipation.

The purpose of this study is to describe the daily pattern of changes in water content in the upper soil layers and to identify the mechanism by which water is added to the soil (deposition or direct absorption). Moreover, the gains in soil water content during the night are compared to the dew amounts recorded by the Hiltner balance, and the losses and gains of water in terms of easily measurable environmental parameters are parameterized.

2. Materials and methods

The measurements were carried out at the Wadi Mashash Experimental Farm in the northern Negev, Israel

TABLE 1. Monthly averages of some meteorological conditions directly affecting dew deposition, together with the same values as measured during the campaigns held in those months, showing that the randomly chosen dates for the field campaigns are representative of the season.

Month		Daytime net radiation ($W\ m^{-2}$)	Nighttime wind speed ($m\ s^{-1}$)	Nighttime air temperature (°C)
Jun	Average	252.94 ± 9.07	2.09 ± 0.52	20.48 ± 1.53
	Campaigns	264.73	2.65	21.30
Jul	Average	247.27 ± 9.33	1.92 ± 0.57	23.00 ± 1.21
	Campaigns	267.33	2.29	24.89
Aug	Average	241.77 ± 14.46	1.83 ± 0.45	23.94 ± 1.43
	Campaigns	240.68	1.91	24.30
Sep	Average	195.29 ± 15.49	1.65 ± 0.34	20.50 ± 1.71
	Campaigns	207.10	1.48	19.44
Oct	Average	141.83 ± 24.83	1.55 ± 0.46	18.64 ± 2.47
	Campaigns	137.08	1.71	17.47

(31°08′N, 34°53′E; 400 m MSL, 60 km from the Mediterranean Sea). Rainfall events occur between October and April, and the mean long-term annual rainfall at the farm is 115 mm. Long-term maximum and minimum temperatures are 14.7° and 4.8°C for January and 32.4° and 18.6°C for July, respectively. Class A pan evaporation is 2500–3000 mm per year. The soil is a sandy loam Aridisol (loess) with 13% clay, 15% silt, and 72% sand, and a porosity of 0.45.

Data were collected during nine 24-h field campaigns that took place during the dry season of 2002. A total of 124 mm of rain was recorded during the rainy season of 2001/02 (previous to the above-mentioned measuring period). The first campaign took place 10 weeks after the last rainfall of the 2001/02 season (29 March, 13.5 mm), and the last one ended about 2 h before the first rainfall of the next season (30 October). The dates of the remaining seven campaigns were randomly spread in between.

Daytime monthly averages of net radiation and nighttime monthly averages of wind speed and air temperature are presented in Table 1. The corresponding values, as measured during the campaigns held in those months, are noted as well, showing that the randomly chosen dates for the field campaigns are representative of the season.

During each campaign, the 100-mm topsoil was sampled hourly, and gravimetric water content (GWC) of the samples was determined at 10-mm increments. A micrometeorological station was installed nearby for continuous measurement of incoming and reflected shortwave radiation with two pyranometers (CM5, Kipp & Zonen[1]); net radiation (Q-7, Campbell Scientific Inc.); wind speed at four levels (2, 1, 0.5, 0.25 m) with cup anemometers (014A, Met-One); dry- and wet-bulb temperatures at 1-m height using a self-designed aspirated psychrometer; and soil heat flux at three different locations in the field with heat flux plates (HFT-3, Campbell Scientific Inc.) installed at a depth of 50 mm, and temperature measurements above them at 10-mm intervals, using differentially wired thermocouples. Data were measured and collected every 10 s and averaged every 30 min by a datalogger (23X, Campbell Scientific Inc.). In addition, the changes in mass of an improved microlysimeter (Ninari and Berliner 2002) (186 mm in diameter and 550 mm of effective depth with an additional 50 mm of polyurethane insulation) were recorded by placing the microlysimeter in a pit and weighing it every half hour. The scale [A&D Engineers, Inc. (AND), maximum weighting capacity of 30 kg] had a resolution of 0.1 g, which resulted in a resolution of 0.004 mm (equivalent depth of water) or 5.11 W m^{-2} (in energy terms). The output of the scale was registered automatically every half hour by a palm computer (48GX, Hewlett Packard). Dew deposition amounts were measured using a conventional Hiltner dew balance (Lambrecht Ltd.) using the original windshield.

Table 2 summarizes the environmental conditions of eight out of the nine campaigns (a datalogger failure occurred during the 27–28 August campaign) for the daytime (from sunrise to sunset) and the nighttime (from sunset to sunrise) separately. Maximum incoming shortwave radiation varied from more than 950 W m^{-2} at the beginning of the measuring period to ~700 W m^{-2} toward the end. Throughout the season, a decrease in both the daytime and nighttime net radiation fluxes is as well apparent. The same trend is evident in the differences between the maximum and the minimum temperatures of both the soil surface and the air. The daytime minimum relative humidity changed significantly during the season, even though daytime averages did not vary much. Maximum relative humidity reached ~95% on all nights. The variance of the average nighttime relative humidity was higher than the corresponding one during daytime. It is worthwhile noticing that the average wind speed on all nights was rather high.

3. Computational procedures

The soil surface temperature was measured at three locations as described in the previous section. On 4 July

[1] Trade or company names are included for the benefit of the reader and do not imply any endorsement or preferential treatment of the product listed by the authors.

TABLE 2. Specifications of the field campaigns' conditions (note the variation between the different days along the dry season).

		Date							
		17–18 Jun	26–27 Jun	18–19 Jul	5–6 Aug	5–6 Sep	25–26 Sep	15–16 Oct	29–30 Oct
Incoming shortwave radiation	Light hours	14	13.5	14	14	12.5	12.5	11.5	11
($W\ m^{-2}$)	Max	991.94	958.15	952.57	944.68	985.16	829.35	744.65	670.49
Net radiation ($W\ m^{-2}$)									
Day	Max	432.74	436.06	426.00	410.68	457.35	367.40	339.24	319.57
	Avg	210.56	198.36	209.25	195.44	195.98	185.96	170.89	126.06
Night	Min	−82.13		−70.54	−74.07	−67.69	−63.82	−55.68	−49.66
	Avg	−53.97		−47.29	−53.07	−55.80	−40.08	−37.91	−39.33
Air temperature (°C)									
Day	Max	35.01	35.91	38.93	37.06	32.72	32.98	33.06	27.56
	Avg	27.07	28.63	32.77	31.03	28.66	27.98	28.84	24.18
Night	Min	15.99		20.42	19.77	16.88	14.95	17.34	12.87
	Avg	20.85		24.98	24.61	20.22	19.22	20.62	16.27
Range		19.03		18.52	17.29	15.84	18.04	15.72	14.68
Soil surface temperature (°C)									
Day	Max	54.40	55.43	55.91	55.78	53.80	47.30	46.09	39.07
	Avg	39.50	39.76	42.32	41.89	40.32	36.07	35.66	29.79
Night	Min	17.05		19.43	20.22	16.61	14.99	16.12	12.66
	Avg	21.55		25.60	25.33	20.48	19.21	20.56	15.99
Range		37.35		36.48	35.56	37.19	32.31	29.98	26.42
Relative humidity (%)									
Day	Min	35.92		44.51	45.51	55.07	57.77	53.47	58.38
	Avg	63.76		61.23	66.60	68.10	65.23	61.23	71.00
Night	Max	96.78		98.79	95.03	93.13	99.32	97.41	96.61
	Avg	86.98		85.44	88.20	85.16	93.71	88.41	85.67
Wind speed ($m\ s^{-1}$)									
Day	Max	7.43		7.03	6.47	6.07	5.91	6.95	5.27
	Avg	3.39		4.08	2.85	2.94	3.73	3.47	2.83
Night	Max	4.623		4.54	4.86	4.86	4.14	5.58	4.22
	Avg	1.95		1.98	1.80	1.74	1.60	2.10	1.49

2001 these average values were compared to soil surface temperature averages obtained using a hand-held infrared thermometer (AG42, Telatemp Corp.) pointed at the areas in which the thermocouples were installed. The analysis of variance of the regression ($r^2 = 0.99$, $p = 0.0000$) indicates that the intercept was not significant and the slope ($=1.07$) was not significantly different from 1. In view of the good agreement between both methods and the fact that it is easier to continuously record thermocouple output, we used the latter to compute average soil surface temperature.

Dewpoint temperature at the soil surface (DPTs) determines if dew will deposit or not. Measurements have not been carried out very close to the soil surface. However, as latent heat fluxes were small and wind speeds relatively high, the vapor pressure gradient was very small, and the dewpoint temperature computed at 1-m height is probably a good rough estimator of the dewpoint temperature close to the soil surface.

The total potential evaporative energy for each day was estimated by summing the half-hourly potential evaporation computed using the Penman–Monteith equation (1) (Monteith 1965), from the time when evaporation began (as derived from microlysimeter measurements, usually at sunrise) to the time when the minimum gravimetric water content was recorded (usually during the early afternoon):

$$\mathrm{PET} = \frac{\Delta(\mathrm{NR} + G) + \rho C_p \dfrac{(e_a^{\mathrm{sat}} - e_a)}{r_a}}{\Delta + \gamma}, \qquad (1)$$

where Δ is the slope of saturated water vapor pressure versus temperature curve (mb C^{-1}); NR is net radiation flux density (W m^{-2}); G is soil heat flux density (W m^{-2}); ρ is dry air density (kg m^{-3}); C_p is specific heat of dry air at constant pressure (J K kg^{-1}); e_a^{sat} and e_a are saturated and ambient water vapor pressure at air temperature (mb), respectively; and r_a is aerodynamic resistance (s m^{-1}) defined as

$$r_a = \frac{\ln(Z_u/Z_0 - \psi_m)\,\ln(Z_e/Z_0 - \psi_v)}{k^2 U_z}, \qquad (2)$$

where Z_U is the height of wind speed measurement; Z_e is the height of water vapor pressure measurement; k is the von Karman constant ($=0.41$); U_z is wind speed at height Z_u (m s^{-1}); ψ_m and ψ_v are integrated diabatic influence functions (Paulson 1970); and Z_0 is the roughness length (m).

The roughness length was computed from wind speed profile measurements obtained under neutral conditions ($-0.01 \leq \mathrm{Ri} \leq 0.01$) and from an additional set of 3D wind speed measurements and was found to be 0.042 $\pm$ 0.074 mm ($n = 1276$).

4. Results and discussion

In order for dew to deposit on a surface, the temperature of the surface should be equal to, or less than, the dewpoint temperature of the air mass it is in contact with. In Fig. 1, the nighttime soil surface temperature and the dewpoint temperature at 1-m height during eight out of the nine campaigns is presented. It can be readily observed that the soil surface temperature did usually not drop below the estimated dewpoint temperature, with the exception of 25–26 September and a short time interval during the early morning of 30 October. From these observations it was to be expected that no dew would deposit on the soil surface. Indeed, dew deposition on the soil surface was not visually observed in any of the campaigns. However, a clear daily cycle in the gravimetric water content of the uppermost 10-mm soil layer was observed, indicating that moisture was absorbed by the soil during the late afternoon and night and evaporated thereafter (henceforth the ‘‘absorption period’’ and ‘‘evaporation period,’’ respectively). A third-degree polynomial was fitted by least squares to the diurnal variation in the measured moisture content of the upper soil layer for each of the nine campaigns. The data and the fitted polynomials are presented in Fig. 2. The polynomial regressions and the coefficients were significant for all campaigns, clearly indicating that the diurnal cycle is not an exceptional occurrence. The minimum water content was recorded close to local standard time (LST), and the maximum around sunrise.

The magnitude of the changes in water content was very small and the average water contents throughout the entire measurement period very low: a maximum of approximately 2% and a minimum that ranged from 1% to 1.5%. These gravimetric water contents correspond to approximately 4×10^2 and 5×10^6 bar, respectively [computed using the Van Genuchten (1980) formulation]. The relative humidity in the soil pores, corresponding to the above-mentioned potentials [computed using the Kelvin equation (Hillel 1971)] reached a maximum of 75% before sunrise and dropped close to 0 at noon. These are conditions for which the dominant mechanism is adsorption, rather than capillary condensation (Philip and de Vries 1957).

Third-degree polynomials were also fitted to each of the remaining 10-mm layers for all campaigns. In Fig. 3, for example, the best-fit lines, drawn using the coefficients of third-degree polynomials, for the uppermost six 10-mm soil layers for 27–28 August are presented. The fact that the amplitude in the water content decreased with depth is evident. This pattern repeated itself for all campaigns, although the depth at which no change was detected varied throughout the season. An objective criterion was needed to determine the depth to which the daily change in water content penetrated. In Table 3 the levels of significance for the coefficients of the polynomials are presented, with the exception of those that correspond to the intercept, as they were always significant but are not relevant to the particular issue under consideration. Layers in which a daily change in water content could be observed were defined as such, if at least two of the three coefficients (intercept

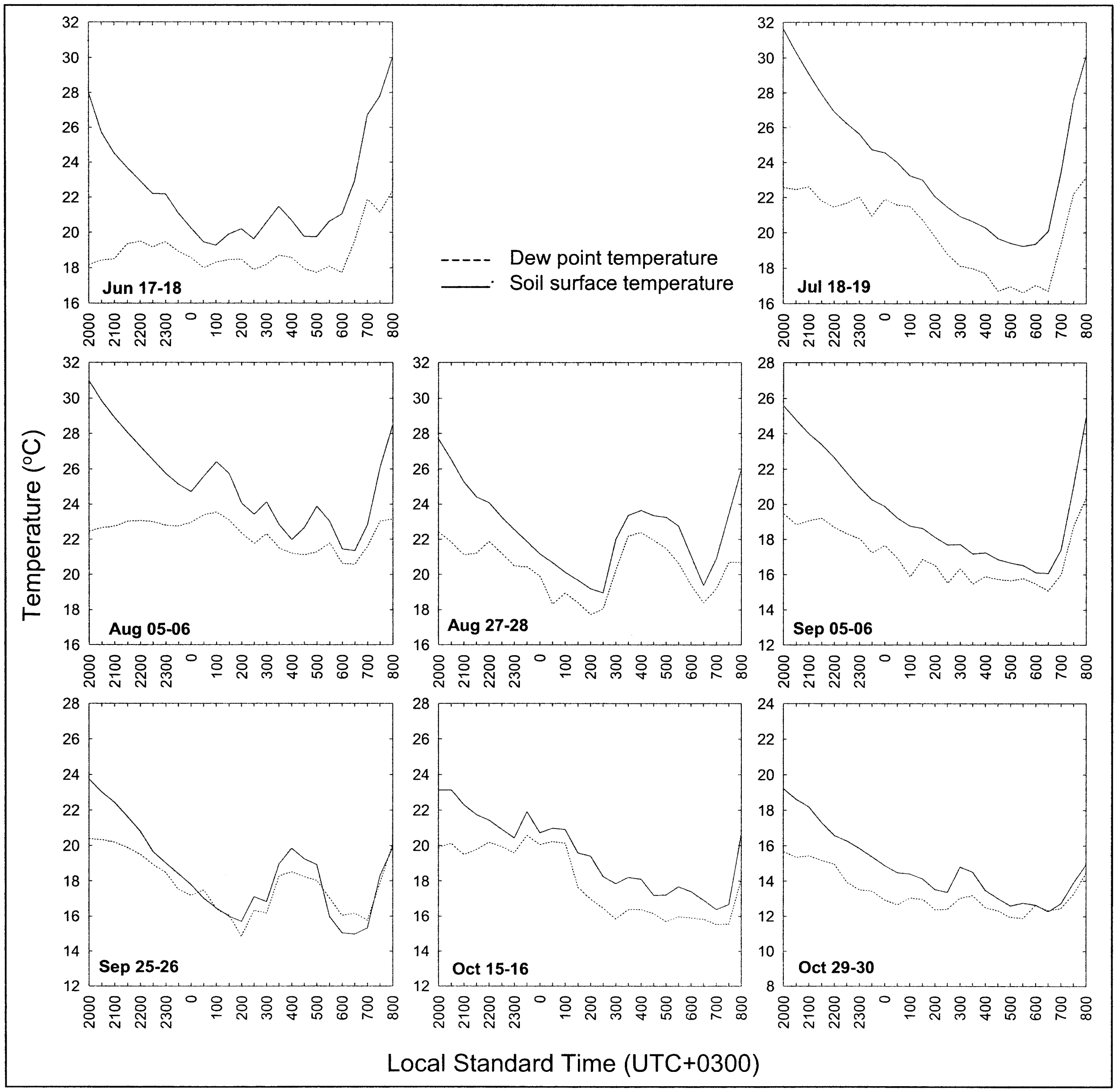

FIG. 1. Nighttime soil surface and dewpoint temperatures during the eight field campaigns. The soil surface temperature did not drop to the dewpoint temperature, with the exception of 25–26 Sep and a brief period during the early morning of 30 Oct.

excluded) were significant. This approach may be illustrated by the best-fit lines presented in Fig. 3 and the corresponding data in Table 3 from which we can conclude, for example, that on 27–28 August the daily change occurred in the five uppermost layers of the soil, and no change in water content was detected below that depth.

Diurnal changes in the water content of the uppermost soil layer have been predicted by various models (e.g., Parlange et al. 1998) that are based on the theories that describe the coupled flow of energy and mass in the soil (e.g., Philip 1957; de Vries 1958; Milly 1982, 1984). The role of water vapor transport was recognized and incorporated in these models. Under extremely dry conditions, as was the case during the present study, water movement in the liquid phase becomes negligible, and the change of water content at any given depth will thus be the result of water vapor movement and physical adsorption or desorption (Scanlon and Milly 1994). Qin et al. (2002) presented a detailed model that linked energy fluxes in the atmospheric boundary layer to the coupled transport of mass and energy in a sand dune in the Negev Desert. Their model does not distinguish between capillary flow due to condensation of dew and vapor adsorption, but it does predict diurnal changes in the water content of the uppermost sand layers that cor-

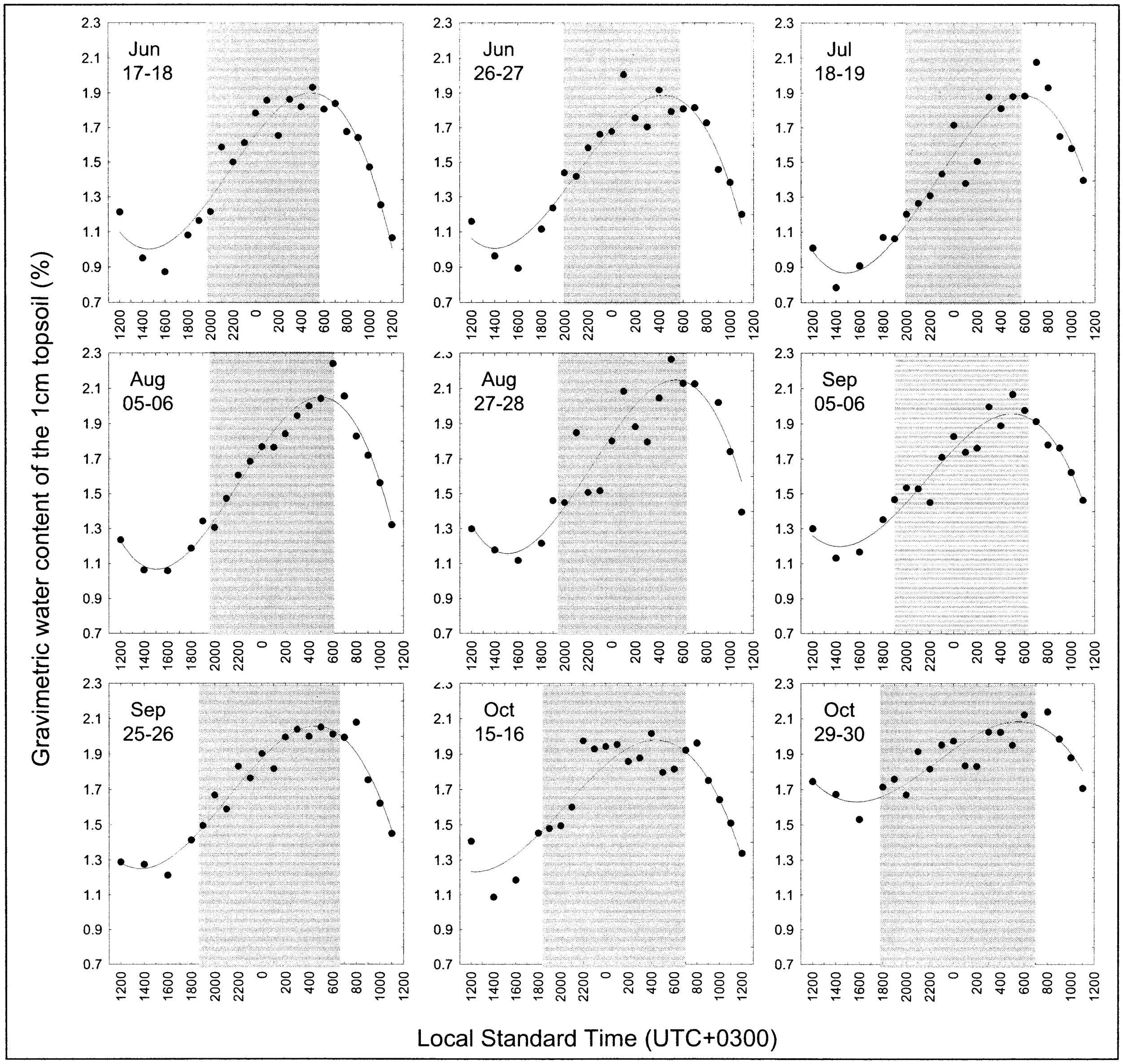

FIG. 2. Diurnal patterns of the gravimetric water content of the uppermost 1-cm soil layer. Gray areas indicate nighttime.

respond qualitatively with our data. No dataset that describes diurnal moisture changes in the upper soil layers for such extremely dry conditions has been previously presented.

The total change in water content (expressed as equivalent depth of water) was computed by summing the changes in volumetric water content for the depths for which a change was defined (as detailed above). Figure 4 presents the adsorption amounts measured with the microlysimeter and those measured with the soil samples. The maximum values on 17–18 July and 27–28 August are slightly higher for the soil samples, while on 26–27 July and 25–26 September the microlysimeter showed slightly higher amounts. No systematic under- or overestimation is evident. The trends are similar and the differences are rather small. The detection of the changes in the microlysimeter's mass, together with the above-mentioned finding that there are no changes in the water content of the deeper layers of the soil, clearly indicates that the addition of water to the soil's uppermost layer is due to absorption of atmospheric water vapor rather than redistribution of water within the soil.

In Fig. 5 the total amounts of added water during the absorption period, as measured using the microlysimeter and the Hiltner dew balance, were plotted against those derived from the total changes in water content of the soil samples. The agreement between the amounts measured with the soil samples and the microlysimeter on one hand and the clear underestimation of the Hiltner balance on the other hand are noteworthy. The average

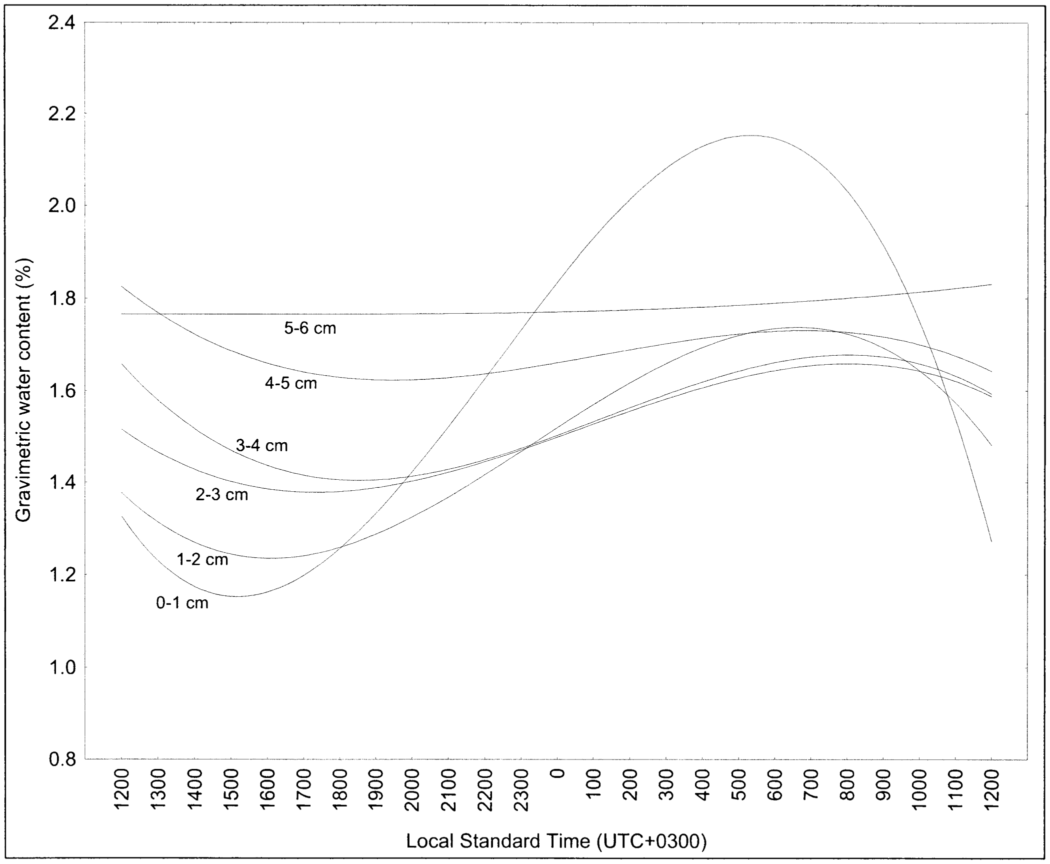

Fig. 3. Best-fit lines obtained from a third-degree polynomial regression analysis for the uppermost six 1-cm soil layers for 27–28 Aug.

water gain per absorption period was 0.26 and 0.25 mm for the soil samples and the microlysimeter, respectively, with corresponding standard deviations (std dev) of 0.05 and 0.04. The average gain detected by the Hiltner balance (for the same dates) was in contrast much lower: 0.08 mm (std dev=0.01). The fact that during the night no signs of dew deposition could be observed on the soil surface indicates that the process by which water was added to the soil was adsorption rather than dew deposition.

A salient feature of Fig. 2 is that the maximum gravimetric water content of the first centimeter of the soil, which for all campaigns reaches a similar value (1.97 ± 0.08%) and is apparently not related to the prevailing environmental conditions. In contrast, the minimum soil water content varies throughout the season and will therefore be the factor that determines the total amount of absorbed water. The seasonal variation of the minimum soil water content suggests that it may be controlled by the atmospheric conditions and therefore should be linked to some index that integrates the various relevant factors. The ''dryness of the atmosphere'' is usually defined as being equal to the maximum rate at which water could be evaporated from a given surface under the prevailing atmospheric conditions (i.e., potential evaporation) and is computed using the Penman equation (Monteith 1965). It was therefore hypothesized that the greater the ''dryness of the atmosphere,'' the lower will be the minimum water content. The period during which the cumulative daily potential evaporation was computed began with the onset of evaporation (as determined from the microlysimeter measurements) and ended at the time at which the minimum water content in the uppermost soil layer was reached. In Fig. 6 the plot of daily accumulated potential evaporation as a function of minimum measured daily water content in the uppermost soil layer is presented. A strong negative linear correlation was found ($r = -0.93$, $p = 0.0007$).

The strong dependence of the minimum water content on the total antecedent potential evaporation together with the previously mentioned fact that a relatively constant maximum water content was observed implies that a correlation between the total antecedent potential evaporation and the amount of water absorbed by the soil during the following ''absorption period'' must exist. In Fig. 7 the total water gain is presented as a function of the antecedent potential evaporation. The correlation is indeed very good ($r = 0.94$, $p = 0.0005$).

For predictive purposes, however, this approach has one serious drawback, namely, that the time of the day

TABLE 3. Significance levels of the slope parameters of third-degree polynomial fit of the GWC vs LST of each of the ten 1-cm layers at the different dates (GWC = $b_0 + b_1\mathrm{LST} + b_2\mathrm{LST}^2 + b_3\mathrm{LST}^3$). Layers in which changes in water content occurred were defined as those for which at least two of the coefficients were significant ($\boldsymbol{\alpha} = 0.05$). The boldface values mark the defined layers.

Date		0–1 cm	1–2 cm	2–3 cm	3–4 cm	4–5 cm	5–6 cm	6–7 cm	7–8 cm	8–9 cm	9–10 cm
17–18 Jun	r^2	**0.93**	**0.92**	**0.79**	**0.57**	**0.62**	0.51	0.51	0.35	0.28	0.39
	b_1	**0.0018**	**0.0000**	**0.0094**	**0.0480**	**0.0353**	0.9191	0.2493	0.9975	0.8248	0.6476
	b_2	**0.0000**	**0.0000**	**0.0010**	**0.0485**	**0.0291**	0.8269	0.5605	0.7222	0.8519	0.7231
	b_3	**0.0000**	**0.0000**	**0.0009**	**0.0694**	**0.0550**	0.8789	0.6736	0.6931	0.7600	0.4231
26–27 Jun	r^2	**0.92**	**0.87**	**0.93**	0.62	0.50	0.27	0.48	0.53	0.60	0.18
	b_1	**0.0166**	**0.0051**	**0.0472**	0.9519	0.9589	0.1242	0.2981	0.0276	0.1241	0.1218
	b_2	**0.0000**	**0.0001**	**0.0038**	0.4421	0.5298	0.2112	0.3865	0.0433	0.3397	0.2928
	b_3	**0.0000**	**0.0000**	**0.0021**	0.4186	0.4530	0.2463	0.4080	0.0498	0.5681	0.3832
18–19 Jul	r^2	**0.92**	**0.84**	**0.70**	0.20	0.47	0.23	0.49	0.21	0.15	0.30
	b_1	**0.0024**	**0.0018**	**0.0522**	0.2847	0.3535	0.4322	0.1384	0.4944	0.7419	0.1386
	b_2	**0.0000**	**0.0001**	**0.0060**	0.2808	0.1426	0.3828	0.0864	0.2934	0.5418	0.1531
	b_3	**0.0000**	**0.0000**	**0.0033**	0.3454	0.1154	0.4396	0.1084	0.2435	0.4937	0.2209
5–6 Aug	r^2	**0.95**	**0.79**	**0.77**	**0.82**	0.37	0.37	0.28	0.35	0.32	0.47
	b_1	**0.0001**	**0.0999**	**0.0305**	**0.0216**	0.3588	0.4199	0.9845	0.7223	0.8631	0.2101
	b_2	**0.0000**	**0.0032**	**0.0029**	**0.0020**	0.1754	0.2853	0.6533	0.9206	0.6996	0.6642
	b_3	**0.0000**	**0.0008**	**0.0018**	**0.0016**	0.1492	0.3055	0.5427	0.8229	0.5445	0.9915
27–28 Aug	r^2	**0.85**	**0.84**	**0.62**	**0.37**	**0.25**	0.07	0.11	0.05	0.01	0.24
	b_1	**0.0065**	**0.0025**	**0.0234**	**0.0389**	**0.0259**	0.9992	0.8612	0.4341	0.8450	0.5224
	b_2	**0.0001**	**0.0001**	**0.0089**	**0.0324**	**0.0299**	0.9806	0.9818	0.4328	0.7920	0.3375
	b_3	**0.0000**	**0.0000**	**0.0107**	**0.0446**	**0.0398**	0.9128	0.9485	0.4302	0.7633	0.3089
5–6 Sep	r^2	**0.93**	**0.71**	**0.69**	**0.58**	0.46	0.55	0.50	0.32	0.27	0.33
	b_1	**0.0073**	**0.0454**	**0.0069**	**0.0042**	0.1861	0.1242	0.5427	0.7264	0.3725	0.3571
	b_2	**0.0000**	**0.0082**	**0.0025**	**0.0082**	0.1463	0.2112	0.6630	0.7328	0.3912	0.3577
	b_3	**0.0000**	**0.0066**	**0.0034**	**0.0241**	0.1980	0.2463	0.5711	0.6172	0.3394	0.4691
25–26 Sep	r^2	**0.95**	**0.81**	**0.55**	**0.51**	0.41	0.52	0.46	0.24	0.37	0.47
	b_1	**0.0157**	**0.2166**	**0.5095**	**0.0006**	0.7977	0.1386	0.3238	0.1131	0.1130	0.0181
	b_2	**0.0000**	**0.0042**	**0.6826**	**0.0004**	0.2805	0.6296	0.8588	0.2245	0.3308	0.0641
	b_3	**0.0000**	**0.0005**	**0.3445**	**0.0004**	0.1254	0.8494	0.6625	0.3720	0.6278	0.1779
15–16 Oct	r^2	**0.83**	**0.50**	0.45	0.24	0.25	0.09	0.13	0.22	0.22	0.05
	b_1	**0.5346**	**0.2997**	0.2452	0.6161	0.2691	0.8890	0.8956	0.6818	0.7741	0.3659
	b_2	**0.0099**	**0.0457**	0.0604	0.3223	0.1441	0.6705	0.6944	0.3847	0.4500	0.3312
	b_3	**0.0007**	**0.0306**	0.0302	0.2301	0.1196	0.5771	0.6410	0.2872	0.2858	0.3280
29–30 Oct	r^2	**0.74**	**0.42**	**0.39**	**0.31**	0.12	0.14	0.08	0.07	0.18	0.48
	b_1	**0.0098**	**0.0693**	**0.0160**	**0.0209**	0.4093	0.9714	0.5832	0.7153	0.4973	0.0054
	b_2	**0.0004**	**0.0253**	**0.0074**	**0.0130**	0.2740	0.9308	0.4901	0.5823	0.3220	0.0233
	b_3	**0.0001**	**0.0203**	**0.0065**	**0.0124**	0.2328	0.9916	0.4834	0.5496	0.2829	0.0555

at which water absorption starts should be known. This information is usually not available. The link between the total water adsorption and the total daylight potential evaporation of the immediately preceding day was therefore tested. The results are presented in Fig. 8, and the correlation is very good and significant ($r = 0.95$, $p = 0.0004$). For a loess soil in the Negev Desert the total water adsorption during the dry period may, therefore, be determined using the following equation:

$$\mathrm{ApN} = 0.09 + 0.04\ \mathrm{PE}, \qquad (3)$$

in which ApN (mm) is the total daily water gain during the ''absorption period'' and PE (mm) is the total potential evaporation for the preceding period (sunrise to sunset) computed using the Penman equation.

Kosmas et al. (1988) concluded that the diurnal fluctuations of soil moisture content due to water vapor adsorption increased with increasing clay content, a result that coincides with the findings of Li (2002). From the foregoing it may be concluded that Eq. (3) is probably site specific. Moreover, the influx of moist air during the late afternoon, which is a characteristic of the coastal desert in which this trial was carried out, will probably also contribute to the site specificity of (3). These facts indicate that equations will have to be derived for different soil–climate combinations, but the approach should be valid for coastal deserts in the southern Mediterranean. Moreover, the specificity is compensated by the fact that the proposed model is much simpler than the one proposed by Kosmas et al. (1988).

5. Summary and conclusions

The results of this study indicate that in the area in which this study was carried out actual dew deposition *on a bare soil surface* is probably a rare occurrence. There is, however, a clear discernible daily cycle of water content in the upper soil layers. The lack of any evidence of soil surface wetting led to the conclusion that the main process responsible for the observed diurnal change in water content is the direct adsorption of water vapor by the soil. The maximum water content attained during each of the nine campaigns was relatively constant and independent of the prevailing me-

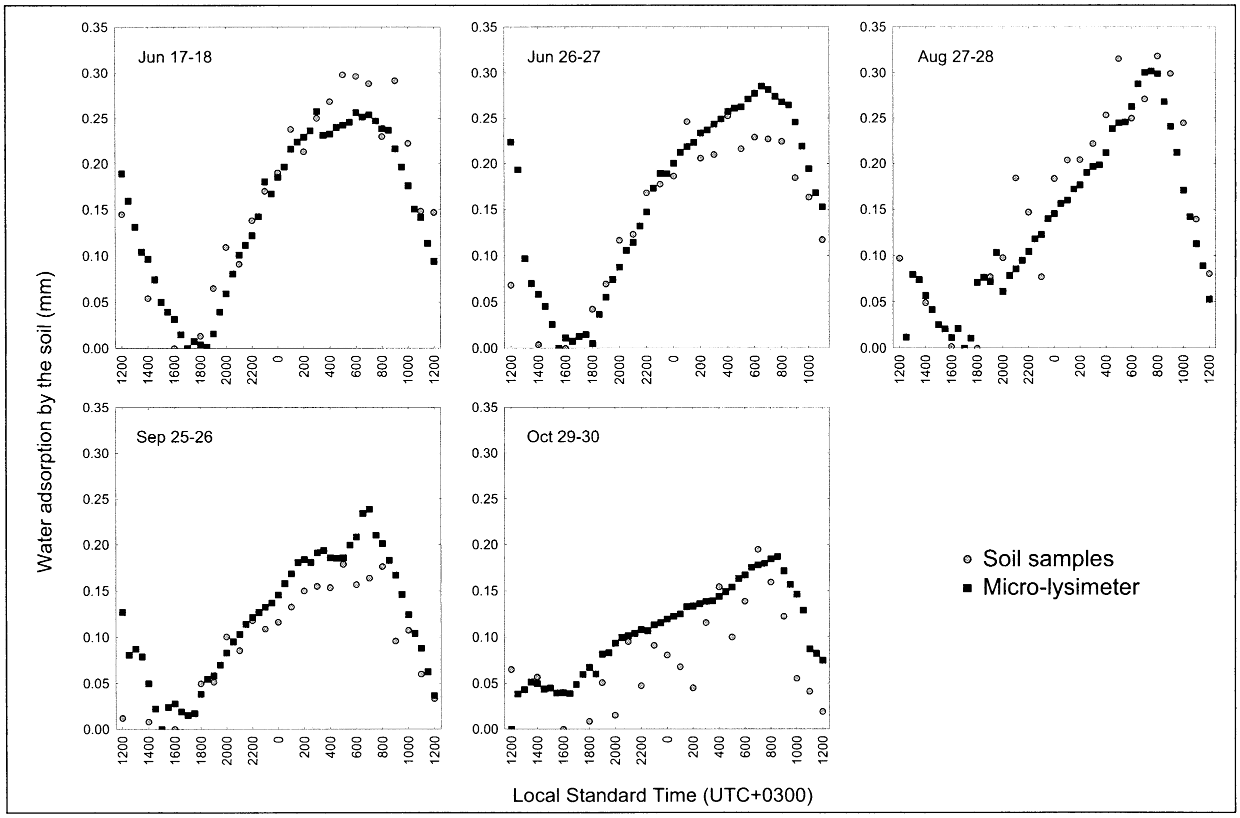

FIG. 4. Diurnal patterns of total water content in the soil (expressed as equivalent water depth) as determined with the microlysimeter and with soil samples for five representative dates. The maximum depth to which computations were carried out for each date is described in the text.

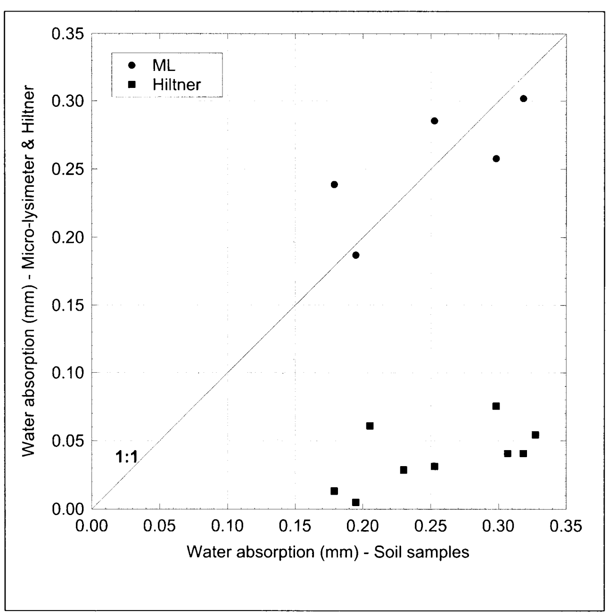

FIG. 5. Total amount of water added to the soil profile (expressed as equivalent water depth) during the night as measured using the microlysimeter and the Hiltner dew balance, as compared to the gains computed from water content changes in the soil samples.

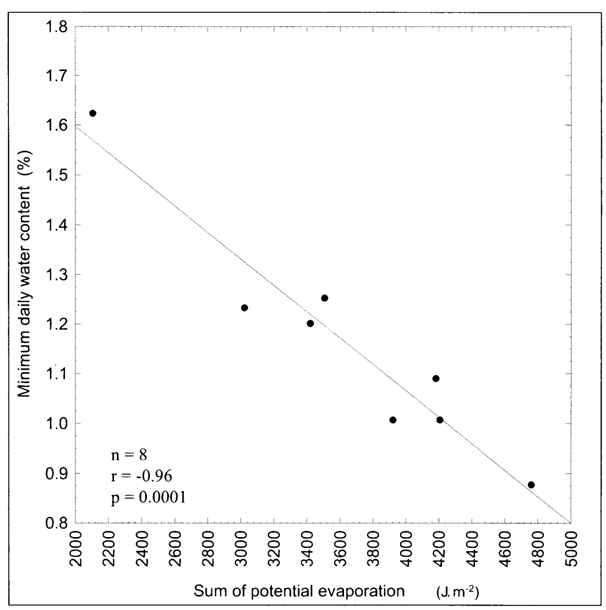

FIG. 6. Minimum daily water content as a function of the sum of potential evaporation from the beginning of the evaporation process until the minimum water content was reached.

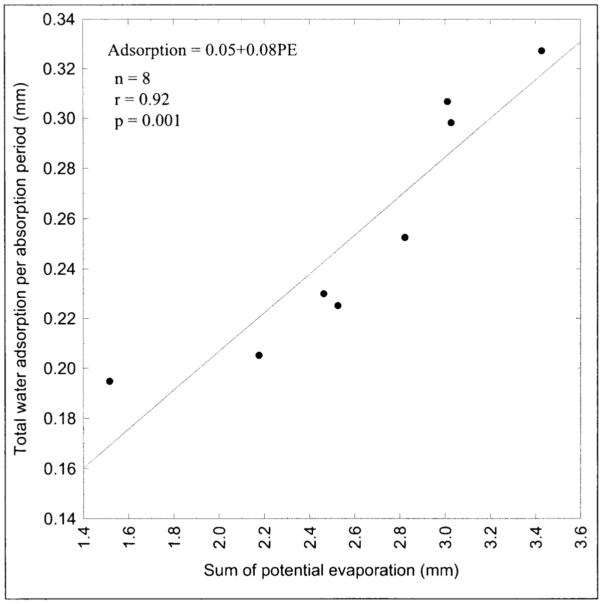

FIG. 7. The dependence of total water adsorption per night (expressed as equivalent water depth) on the sum of potential evaporation (PE) from the time when evaporation began to the time at which the minimum water content was reached on the day previous to the period during which adsorption takes place.

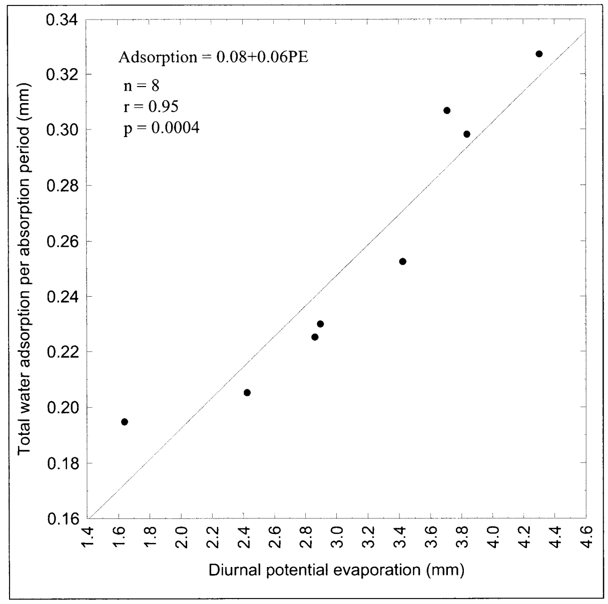

FIG. 8. The dependence of total water adsorption (expressed as equivalent water depth) on the diurnal PE (total PE from sunrise to sunset on the day previous to the night during which adsorption was monitored).

teorological conditions. Therefore, the minimum water content that was reached during the preceding day determines the total amount of water vapor adsorbed. A strong and significant correlation between the total adsorption of water vapor by the soil during the ''absorption period'' and the total potential evaporation between sunrise and sunset of the previous day was found. Based on this finding, an empirical model based on commonly available data is proposed in order to predict the total amount of water adsorbed by the soil during the absorption period. The proposed model is probably site specific but very simple and easy to implement.

The fact that there is no evidence of the deposition of liquid water on the soil surface has two additional consequences. The first is that even though the magnitude of the latent heat fluxes are larger than those previously reported, the long-held belief that dew is of prime importance in this type of ecosystem needs to be reassessed. The second consequence is the possible effects on remote sensing techniques. The reported changes in some of the remotely sensed data should be related to the small changes in the soil surface water content, as it appears that no water drops are formed on the surface of the soil.

An additional and interesting consequence of the finding of this study is that in coastal deserts there is, during the dry season, a daily cycle of water vapor exchange between the soil and the atmosphere. The latent heat flux density during the early morning may reach 20% of the net radiation flux density and decreases thereafter. The magnitude of this flux is of the same order of magnitude as the expected soil heat flux density (Stull 1988) and should therefore be considered when energy and water balances at the soil surface are computed for these regions.

Last, it can be stated that, in the area in which this study was carried out, artificial condensing plates of any kind cannot be used to evaluate dew deposition quantities on a bare soil surface.

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