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TEMPERATURE MODELLING OF WET AND DRY DESERT SOILS

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SUMMARY

(1) As part of a study on thermal relations of desert plants, soil temperatures at various depths, soil evaporative water loss, radiation components, wind speeds and air temperatures were measured hourly over a 24-h period for a naturally dry and an artificially wet soil in the north-western Sonoran Desert in California. Also, the thermal conductivity, volumetric heat capacity and water potential of the soil were determined in the laboratory for a wide range of soil-water contents. A computer model was then developed to predict soil temperature at various depths.

(2) The measured soil temperature was generally within 1 °C of that predicted by the model, and the measured net energy exchange of the soil with the atmosphere closely agreed with the predicted heat storage in the soil. A sensitivity analysis showed that maximum soil surface temperatures were markedly influenced by shortwave radiation, wind speed and air temperature.

(3) Model simulations showed that nearly all of the decrease in maximum temperature of a wet soil occurred above a soil-water potential of −0·2 MPa. The damping depth for daily temperature changes was about 10 cm for both the wet and the dry soils.

(4) When the soil was shaded by a common bunchgrass, which influenced shortwave and longwave radiation as well as wind speed profiles, simulated maximum soil surface temperatures decreased about 2 °C for every 10% increase in shading, indicating that nurse plants can have a substantial effect on the local microclimate.

INTRODUCTION

Soil surface temperatures in deserts can exceed 70 °C and have even been reported to reach 80 °C (Hadley 1970; Körner & Cochrane 1983). Small plants in general and seedlings in particular can be strongly affected by such high soil temperatures due to their proximity to the soil surface and their relatively low thermal capacity. Thermal properties of the soil also influence mineral uptake as well as growth and development of roots. Many thermal and other properties of soil are markedly affected by its water content, considered here under the extremes of wet and dry conditions.

Heat movement in soils usually is mainly by conduction (van Wijk 1966; de Vries 1975; Nobel 1983):

$$J_{\mathrm{H}}^{\mathrm{C}} = -K^{\mathrm{soil}} \frac{\delta T}{\delta z} \qquad (1)$$

where $J_{\mathrm{H}}^{\mathrm{C}}$ is the heat flux density by conduction, K^{soil} is the thermal conductivity coefficient of the soil, and $\delta T/\delta z$ is the temperature gradient.* K^{soil} varies with the water

* *Principal symbols*: *a*, shortwave absorptance; $C_{\mathrm{P}}^{\mathrm{soil}}$, soil volumetric heat capacity; *d*, damping depth; e_{IR}, longwave emittance; $\mathrm{IR_d}$, longwave radiation downward from sky; $\mathrm{IR_u}$, longwave radiation upward from ground; $J_{\mathrm{H}}^{\mathrm{C}}$, heat flux density by conduction; $J_{\mathrm{H}}^{\mathrm{L}}$, latent heat flux density; J_{wv}, water vapour flux density; K^{soil}, soil thermal conductivity coefficient; S, shortwave radiation; T_z, temperature at height *z*; u_z, wind speed at height *z*; ψ^{soil}, soil water potential.

content of the soil, as does the volumetric heat capacity, C_P^{soil}. These two variables can be used to calculate the damping depth (d) where the soil temperature variation decreases to 1/e (37%) of the value at the soil surface:

$$d = \left(\frac{p\,K^{soil}}{\pi\,C_P^{soil}}\right)^{\frac{1}{2}} \tag{2}$$

with p being the time period of interest, usually a day or a year (Nobel 1983).

The present study characterized the soil thermal environment at a site in the northwestern Sonoran Desert where physiological aspects of the vegetation have been extensively investigated (e.g. Nobel 1976, 1977, 1981). The dominant species at the site are the bunchgrass *Hilaria rigida* (Thurb.) Benth. ex Scribn. and the leaf succulent *Agave deserti* Engelm., with other succulents such as *Ferocactus acanthodes* (Lem.) Britton & Rose var. *acanthodes* also being common. These three species have shallow roots, with mean depths of only 8–11 cm (Nobel 1976, 1977, 1981), and hence attention was focused on the upper part of the soil. Various environmental variables and soil properties were measured so that an energy budget for the surface and subsurface soil layers could be calculated. A computer model was then developed, allowing predictions of soil temperatures under other environmental conditions.

MATERIALS AND METHODS

Field site

Soil temperatures and various energy budget variables were measured over 24 h on 10 March 1984 at the University of California Philip L. Boyd Deep Canyon Desert Research Center (at 33°38′N, 116°24′W, 850 m elevation), about 8 km south of Palm Desert, California. The dry soil occurred naturally; its soil water potential was less than −6 MPa in the upper 20 cm and was −4·8 MPa at 40 cm, and its volumetric soil water content in the upper 50 cm was 3%. Soil water potential (ψ^{soil}) was measured with Wescor PT 51-05 soil thermocouple psychrometers, and volumetric water content was determined gravimetrically (Young & Nobel 1986). To prepare the wet soil, a series of narrow holes was made and kept filled with water in the morning of the day before measurements; the soil was re-wetted every 3–5 hours to maintain the volumetric soil water content at 29 ± 3%. Both sites were level, unshaded, bare patches of ground approximately 1·0 m in diameter in a region where the vegetation was sparse and less than 0·3 m tall for at least 10 m from the sites considered. Samples of the soil, which is formed from disintegrating granite (Nobel 1976), were collected for measurements of thermal properties.

Environmental variables

Soil temperatures were measured at the centres of the wet and dry sites with copper-constantan thermocouples 0·51 mm in diameter placed at ground level and at depths of 0·5, 3, 9, 29 and 44 cm (the last being the depth of the deepest roots for *A. deserti*, *F. acanthodes* and *H. rigida*). For measurements of air temperatures, shielded copper-constantan thermocouples 0·13 mm in diameter were placed at 0·003, 0·01, 0·1, 0·3, 0·5, 1·0 and 2·0 m above the soil surface. Air and soil temperatures represent the average of five readings at 2-min intervals centred on the indicated times.

Direct plus diffuse shortwave solar radiation (S) was measured with a Moll–Gorczynski solarimeter and longwave (infra-red) radiation with an Eppley PIR hemispherical pyrgeometer, facing upward at ground level for the downward sky radiation (IR_d) and facing downward 0·2 m above the ground for longwave radiation emanating from the soil (IR_u). The net radiation balance thus is

$$\text{net radiation flux density} = aS + a_{IR}IR_d - IR_u \tag{3}$$

where a is the shortwave absorptance of the ground (equal to $1-r$, where r is the ground reflectance) and a_{IR} is its longwave absorptance, which is equal to its longwave emittance e_{IR} (Watts 1975; Nobel 1983). Thus a_{IR} could be calculated from $a_{IR} = e_{IR} = IR_u/\sigma\ T_o^4$, where σ is the Stefan–Boltzmann constant and T_o is the soil surface temperature, here in kelvin units (Nobel 1983).

The flux density of water vapour (J_{wv}) from the soil was measured with a LiCor-1600 steady-state porometer. A LiCor LI-1600-02A cylindrical chamber with the cover removed was placed for about 2 min each hour on a 6 cm × 5 cm rectangular plastic frame inserted 2 cm into the ground to minimize lateral movement of soil air. The latent heat flux density due to evaporation (J_H^L) could thus be represented by

$$J_H^L = J_{wv}\ H_{vap} \tag{4}$$

where H_{vap} is the heat of vaporization of water (2·44 MJ kg^{-1} at 25 °C).

Wind speed was measured at heights of 0·003, 0·01, 0·03 and 0·1 m with hot-bead anemometers (Carr 1978) calibrated individually with a Lambrecht 641N hot-wire anemometer and at 0·3, 0·5, 1·0 and 2·0 m with Thornthwaite 106 cup anemometers. Wind speeds at height z (u_z) were averaged for 1-h intervals centred on the points indicated and equated to $(u^*/\mathrm{k})\ln[(z-\delta)/z_o]$, where u^* is the friction velocity, k is von Karman's constant (0·41), δ is the zero plane displacement, and z_o is the roughness length (Monteith 1973; Campbell 1977). The sensible heat flux density convected in the air above the ground (J_H^C) then is

$$J_H^C = C_P^{air} \frac{\mathrm{k}u^*}{\ln\left(\dfrac{z+z_o}{z_o}\right)} (T_z - T_o) \tag{5}$$

where C_P^{air} is the volumetric heat capacity of air (1200 J m^{-3} K^{-1} at 25 °C), T_z is the temperature at height z above the ground, and T_o is the temperature of the soil surface (Monteith 1973; Campbell 1977; Grace 1981).

Soil physical properties

The soil thermal conductivity coefficient K^{soil} (Eq. 1) was determined in the laboratory using soil from the field site compacted to the bulk density measured in the field (1520 kg m^{-3}; Blake 1965). A 1000-W hot-plate was used to heat one side of soil slabs of various water contents. Steady-state measurements of heat flux density were made with Thermonetics H11-18-3-G heat flux plates, and 0·51-mm-diameter copper-constantan thermocouples were used to obtain the temperature gradient; the effect of the heat flux plates on heat flow was corrected for by the method of Philip (1961).

The soil volumetric heat capacity (C_P^{soil}) is needed to calculate heat storage in volume V of the soil:

$$\text{rate of heat storage} = C_P^{soil}\ V \frac{\Delta T}{\Delta t} \tag{6}$$

where ΔT is the temperature change in time Δt (Lewis & Nobel 1977; Nobel 1983). To calculate C_P^{soil}, known volumes of water at specific temperatures were mixed adiabatically in vacuum-insulated containers with known volumes of oven-dried soil at specific temperatures and the final temperature determined (heat absorption by the container was corrected for). Soil hydraulic conductivity and ψ^{soil} were related to soil volumetric water content using data for the field soil (Young & Nobel 1986).

Computer model

A computer model was developed to predict soil temperatures at various depths as a function of soil physical properties and environmental conditions. The model, which was based on the one developed by Lewis & Nobel (1977), incorporated hourly values of S, IR_d, J_{wv}, $u_{2\ m}$, and $T_{2\ m}$ (the subscript 2 m refers to a height of 2 m above the ground). To minimize problems of advection and atmospheric instability, the measured wind speed and air temperature profiles were used to derive empirical relations between J_H^C (Eq. 5) and $u_{2\ m}$ and $T_{2\ m}$ for various intervals near the soil surface. Thus, besides soil properties such as C_P^{soil} and K^{soil}, both of which vary with soil water content, and a, the model used as input variables the radiation quantities S and IR_d (which themselves could be stimulated), $u_{2\ m}$ and $T_{2\ m}$, which are generally measured in standard weather stations, and J_{wv}, which was measured by the new technique described above. Simulations were done for the two extremes: a dry soil with a low J_{wv} and a wet soil with a high J_{wv}.

For the model, the soil was divided into fifty isothermal layers 1 cm thick plus a surface layer of zero thickness (little change in soil temperature at a depth of 50 cm would be expected over a 24-h period). The energy budget for the surface layer included net radiation (Eq. 3), latent heat loss (Eq. 4), as well as sensible heat conduction into the ground (Eq. 1) and to the air (Eq. 5). Each subterranean layer had heat conduction to or from the two adjacent layers as well as heat storage (C_P^{soil}) times the rate of change of temperature times the layer volume (Eq. 6). After doing an energy budget calculation for each layer, the entire ensemble of fifty-one layers was allowed to interact in a series of iterative steps (Kreith 1973) until temperature convergence to within 0·1 °C of the previous iteration was achieved for each layer (Lewis & Nobel 1977; Nobel 1978).

The effects of nurse plants on soil temperatures were determined by incorporating the effect of the bunchgrass *Hilaria rigida*, a common nurse plant for seedlings of *Agave deserti* and *Ferocactus acanthodes*, on shortwave and longwave radiation reaching the soil surface and on convective heat exchange. A wind speed profile was determined within and above the canopies for a series of 40-cm-tall plants that decreased the shortwave radiation reaching the soil surface by up to 90%, depending on the number of culms per unit ground area. An empirical relationship between soil surface shading and the decrease in wind speed within the canopy was then used to adjust the convective heat exchange between the soil and air in the model, e.g. when a particular plant of *H. rigida* reduced the shortwave radiation reaching the soil surface by 55%, the average wind speed within the canopy was reduced 70% compared with the wind speed at the same heights for an adjacent exposed location.

RESULTS

Environmental variables

On the relatively clear spring day considered, shortwave radiation (S) reached a maximum of about 840 W m^{-2} near solar noon and was effectively zero at solar times of

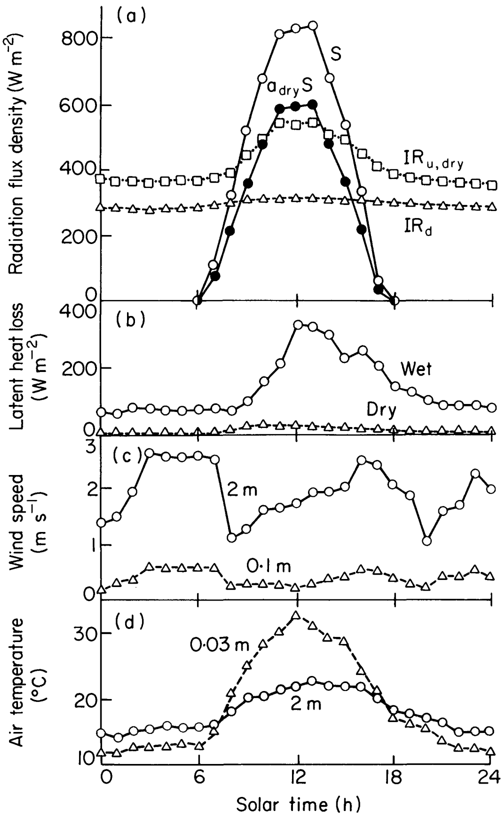

FIG. 1. Daily time courses of (a) incident (S) and absorbed shortwave radiation for the dry soil (a_{dry} S) as well as downward (IR_d) and upward longwave radiation at the surface of a dry soil ($IR_{u,\text{dry}}$); (b) latent heat loss from wet or dry soils; (c) wind speed at the indicated heights above the dry soil; and (d) air temperature above the dry soil.

6 h and 18 h (Fig. 1a). The soil shortwave absorptance (a) averaged 0·71 for the dry soil and 0·76 for the wet one. When the sun was at a low angle in the sky, a tended to be lower due to specular reflection; e.g., at a solar time of 7 h, a was 0·64 for the dry soil and 0·71 for the wet one. The longwave radiation incident on the soil surface (IR_d) averaged 300 W m^{-2} (Fig. 1a) and was approximately 10% higher at midday than in the middle of the night because of the increased temperature of the lower atmosphere and surrounding vegetation at midday. The measured T_o and IR_u indicated that e_{IR} was $0{\cdot}96 \pm 0{\cdot}02$ (mean $\pm$ standard deviation for $n = 6$) for the dry soil and $0{\cdot}97 \pm 0{\cdot}02$ ($n = 6$) for the wet one.

The water vapour flux density was considerably greater from the wet soil than the dry one, and so the wet soil had the greater latent heat loss (Fig. 1b). Over the 24-h period, the water leaving the wet soil corresponded to a depth of 4·94 mm compared with 0·36 mm from the dry soil. Considerably more of the daily evapotranspiration occurred in the afternoon than in the morning for the wet soil, while contributions from these two periods were about equal for the dry soil (Fig. 1b).

The wind speed decreased toward the soil surface (Fig. 1c). Over the 24-h period u_z averaged 1·96 m s^{-1} at 2·0 m above the ground, 1·27 m s^{-1} at 0·5 m, 0·38 m s^{-1} at 0·1 m,

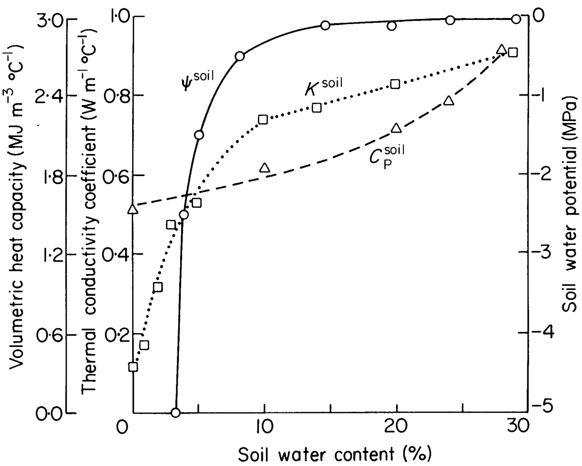

FIG. 2. Variation in soil physical properties with water content. All measurements were made in triplicate using soil from the field site. Data on ψ^{soil} are from Young & Nobel (1986).

and 0·12 m s^{-1} at 0·01 m. Because the artificially wetted site was relatively small, the same wind speed profile was assumed to occur above both the wet and the dry soils. For the bare soil considered, δ averaged 0·5 mm, z_0 averaged 5·7 mm, and u^* averaged 0·11 m s^{-1}.

The greatest daily variation in air temperature occurred closest to the ground. For the dry soil the daily variation (daytime maximum minus night-time minimum) was 28·8 °C at 0·003 m above the ground, 20·1 °C at 0·03 m, 13·3 °C at 0·3 m, and 8·8 °C at 2 m (Fig. 1d). Air temperatures at the ground level were equal to soil surface temperatures, varying 36·6 °C during the 24-h period for the dry soil. Daily variations in air temperatures above the wet soil were less, e.g. 21·2 °C at the ground, 19·6 °C at 0·003 m above the ground, and 16·8 °C at 0·03 m.

Soil physical properties

The soil water potential (ψ^{soil}), thermal conductivity (K^{soil}), and volumetric heat capacity (C_P^{soil}) all increased with increasing soil water content (Fig. 2). Field capacity occurred at about 29% water by volume, which corresponded to a ψ^{soil} of $-0·01$ MPa. A ψ^{soil} of $-1·5$ MPa, which is often used to designate the permanent wilting point of crop plants (Larcher 1980; Nobel 1983), occurred at about 5% water. Increasing the water content between these two limits caused K^{soil} to increase 63% and C_P^{soil} to increase 48%. Although C_P^{soil} did not decrease much below a soil water content of 5%, K^{soil} was quite sensitive to water contents from 0 to 5% (Fig. 2).

Model validation

The measured maximum soil surface temperature was just over 45 °C for the dry soil (Fig. 3a) and just under 30 °C for the wet one (Fig. 3b). At greater depths maximum temperatures occurred later in the day, e.g. for the dry soil the maximum temperature at 9 cm occurred at 16 h compared with 13 h at the surface. Very little change during the 24-h period occurred at a depth of 44 cm, the soil temperature remaining at 16·8 ± 0·5 °C for the dry soil and 15·7 ± 0·8 °C for the wet soil (Fig. 3a, b). Also, hourly temperatures measured in the four cardinal directions 20 cm radially outward from the centre of the dry

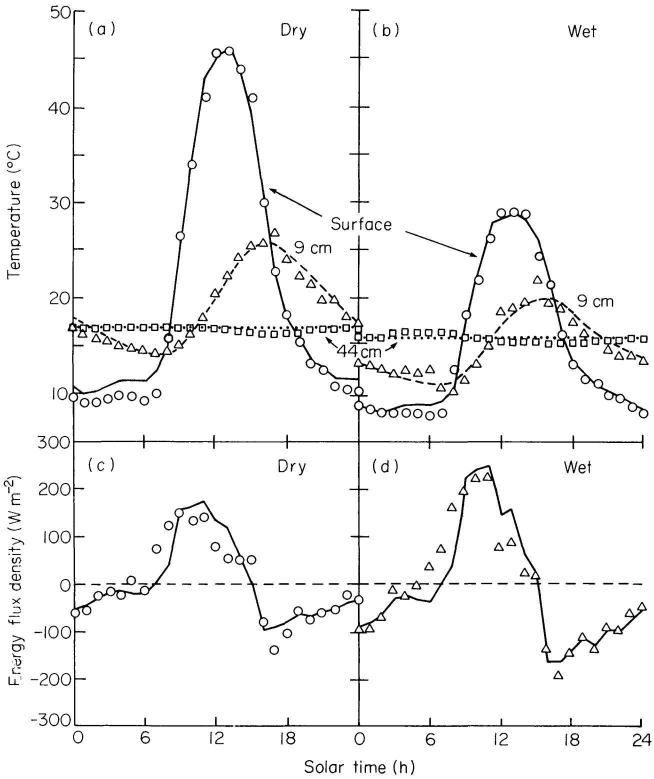

FIG. 3. Predicted (lines) and measured (symbols) soil temperatures at the indicated depths for dry (a) and wet (b) soils. Also, comparison of measured net energy exchange of soil with the atmosphere (symbols) with predicted heat storage in the upper 50 cm of soil (lines) for dry (c) and wet (d) soils.

or wet sites for the surface and depths of 9 and 44 cm generally agreed within 0·5 °C with those at the centre, indicating that the sizes of the wet and dry sites were large enough to warrant using a one-dimensional soil model.

Predictions by the model agreed well with the measured values of soil temperature at all depths, three of which are shown in Fig. 3a. Model predictions were always within 2 °C of the measured values and usually within 1 °C. Of particular importance for the study of high-temperature tolerance of plants, the predicted maximum soil temperatures were within 0·5 °C of the measured maxima. Model predictions for heat storage within the soil were also compared with the net energy exchange at the soil surface, calculated from the three energy terms determined in the field (net radiation − latent heat loss + sensible heat exchange). Except for an apparent lag in the daytime, measured net energy exchange with the environment agreed well with the heat storage in the soil predicted by the model (Fig. 3c, d).

Sensitivity analysis

A sensitivity analysis was performed to determine the influence of various soil and environmental variables on soil temperature, with only changes in the maximum

TABLE 1. Sensitivity analysis of a computer model of temperature changes in desert soil. Soils and microclimatic variables were independently changed as indicated (e.g. $\times \frac{1}{2}$ means halved from the field values), and the effect on maximum soil temperature at the surface (0 cm) and at a depth of 9 cm determined for dry and wet soils.

		Change in maximum temperature (°C)			
		Dry soil		Wet soil	
Variable	Simulated change	0 cm	9 cm	0 cm	9 cm
u_{2m}	$\times \frac{1}{2}$	+8·1	+2·8	+2·0	+0·7
	$\times 2$	−8·8	−3·1	−1·2	+0·1
T_{2m}	−10 °C	−5·5	−4·3	−5·0	−3·9
	+10 °C	+5·4	+4·1	+4·9	+3·7
T^{sky}	+10 °C	+3·0	+2·0	+2·5	+1·7
S	$\times \frac{1}{2}$	−16·5	−6·6	−12·0	−6·4
a	−0·1	−4·5	−1·7	−3·5	−1·9
	+0·1	+4·8	+2·1	+3·7	+1·9
J_H^L	$\times \frac{1}{2}$	+0·4	+0·3	+6·1	+3·4
	$\times 2$	−1·4	−0·7	−8·7	−6·1
K^{soil}	$\times \frac{1}{2}$	+2·0	−3·5	+2·9	−1·8
C_P^{soil}	$\times \frac{1}{2}$	+2·0	+5·4	+2·4	+3·7

temperature being presented here. Halving and doubling u_{2m} changed the maximum surface temperature for the dry soil by +8·1 °C and −8·8 °C, respectively, but for the wet soil these values were only +2·0 °C and −1·2 °C (Table 1). Decreasing and increasing the air temperature at 2 m by 10 °C changed the maximum surface temperature of dry soil by −5·5 °C and +5·4 °C, respectively, with the change at a depth of 9 cm being about 1 °C less (wet soil changes were about 0·5 °C lower than their dry soil counterparts). Smaller changes in soil temperature accompanied 10 °C changes in the effective longwave temperature of the sky (Table 1). Halving the incoming radiation decreased the maximum surface temperatures by 16·5 °C for the dry soil and by 12·0 °C for the wet soil (changes in a had analogous effects). Although halving and doubling the latent heat loss had little effect on dry soil, wet soil surface temperatures were changed by +6·1 °C and −8·7 °C, respectively. Halving K^{soil} raised the surface temperature but lowered the temperature at 9 cm, while halving C_P^{soil} raised both temperatures (Table 1).

Applications of model

Changes in C_P^{soil} without concurrent changes in K^{soil}, such as were done in the sensitivity analysis, would probably not occur naturally, because both variables depend upon soil moisture (Fig. 2). Thus, to determine the influence of soil water potential on daily maximum soil temperatures, K^{soil} and C_P^{soil} as well as a and J_H^L were adjusted in the model to reflect soils of different water content. For the upper layers, increases in ψ^{soil} always decreased maximum soil temperature, while for the lower layers maximum temperatures occurred at an intermediate ψ^{soil} (Fig. 4). The increase in temperature with ψ^{soil} at greater depths primarily reflected the increase in K^{soil} as the soil was wetted. Regardless of depth, however, maximum soil temperatures changed relatively little as ψ^{soil} increased until a ψ^{soil} of −0·5 MPa, where further increases in ψ^{soil} resulted in large declines in soil temperature (Fig. 4).

Shading of the soil surface by the bunchgrass *H. rigida* decreased simulated maximum temperatures of dry and wet soils (Fig. 5). For example, surface temperatures for soil with 50% shading were about 11 °C less than those of unshaded soil for both dry and wet

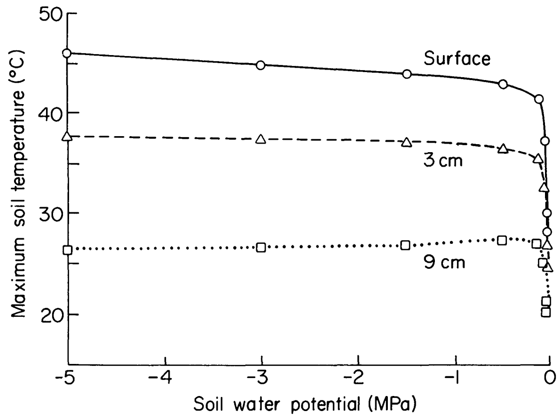

FIG. 4. Influence of simulated changes in ψ^{soil} on daily maximum soil temperatures at the indicated depths. K^{soil} and C_P^{soil} were obtained from Fig. 2; a was determined by interpolation based on the measured absorptances for dry and wet soils and the volumetric soil water content; J_H^L was determined similarly, except that the interpolation was based on the soil hydraulic conductivity (Young & Nobel 1986), a variable controlling evaporative flux density that varies non-linearly with soil water content.

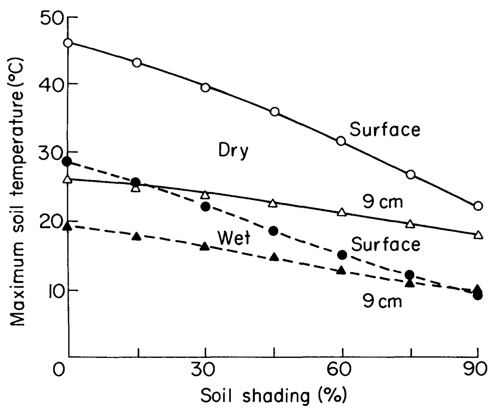

FIG. 5. Predicted effect of shading by a bunchgrass on maximum temperatures at the surface and at a depth of 9 cm for dry or wet soils.

conditions. Similarly, at a depth of 9 cm, soil shaded 50% was 4 °C to 5 °C cooler than for the unshaded condition (Fig. 5). Decreases in shortwave radiation with increased shading proved to be more important than increases in downward longwave radiation and decreased convection due to lower wind speeds, consistent with the sensitivity analysis (Table 1).

DISCUSSION

The physical properties measured here for soil from the north-western Sonoran Desert are consistent with values determined previously for other soils. For soils of various water contents, the soil volumetric heat capacity (C_P^{soil}) averages 1·9 MJ m^{-3} $°C^{-1}$ (de Vries

1966, 1975), which is similar to the value found here at a water content of 14%. C_P^{soil} increased with water content (Fig. 2), as would be expected because the volumetric heat capacity of water is about three orders of magnitude greater than that of the air it replaces. This replacement of soil air by water also increases the soil thermal conductivity coefficient (K^{soil}). For sandy and loamy soils, K^{soil} increased about four-fold as volumetric water content increases from 1% to 10% (de Vries 1975), similar to the present results. Measured values for K^{soil} were between those for loam and sand (de Vries 1975), as would be expected for the field soil, which is on the borderline between the textural categories of loamy sand and sandy loam (Nobel 1976). Similarly, the volumetric water contents found here at field capacity and $-1{\cdot}5$ MPa were intermediate between values for loam and sand (Larcher 1980).

Because both C_P^{soil} and K^{soil} increased with volumetric water content, the damping depth (d, Eq. 2), which depends on their ratio (Eq. 2), remained fairly constant. The increase in K^{soil} with water content leads to greater heat fluxes down into the soil, but more heat is then required to change the soil temperature. For a dry soil with 5% water by volume, d for daily changes is 9·7 cm compared with 9·9 cm for a wet soil with 25% water. The higher C_P^{soil} for wet soils slows their heating, and the potentially greater latent heat loss can suppress the maximum temperature and lead to a lower minimum temperature than for dry soils. The changes in the water flux density from the wet soil over the course of a day (Fig. 1b) paralleled the changes in the water vapour concentration drop from a saturated surface at the soil surface temperature to the air, where the concentration remained at about 3 g m^{-3}. The dry soil tended to lose water more readily in the late morning than in the early afternoon, possibly indicating a decrease in the hydraulic conductivity in the upper part of the soil caused by water depletion.

Daily variations in shortwave radiation established the periodicity in soil temperature fluctuations. Even here, some differences occurred between wet and dry soils. Specifically, the shortwave absorptance was about 10% higher for the wet soil, as observed previously (van Wijk & Scholte Ubing 1966), values being similar to those for sand-dunes (Stanhill 1970). The lower absorptance at lower sun angles is also consistent with previous observations (Watts 1975). The longwave absorptance and emittance of 0·96 and 0·97 measured here for dry and wet soils, respectively, agree with previous measurements on bare soils (Watts 1975).

The computer model accurately predicted soil temperatures for both wet and dry soils (Fig. 3a). The largest discrepancies between measured and predicted values for the wet soil occurred just after the soil was rewetted, suggesting some thermal disturbance then. Other models have also predicted soil temperatures in deserts (e.g. Meikle & Treadway 1979; Mitchell *et al.* 1975). A model similar to that developed here, except that water evaporation was ignored and K^{soil} and C_P^{soil} were assumed constant, predicted air and soil temperatures within about 2 °C at three desert sites (Porter *et al.* 1973; Mitchell *et al.* 1975). At one of their sites, considerable precipitation occurred the night before the test period and, although their model did not include the effect of such added moisture, close agreement occurred between the measured and the predicted values. This may have in part resulted from rapid decreases in soil hydraulic conductivity, and hence latent heat loss, as the soil surface began to dry (cf. Fig. 4).

The sensitivity analysis indicated that decreases in solar radiation, such as might be caused by clouds, nurse plants and micro- and macro-topographic obstructions, resulted in large decreases in soil temperature (Table 1). Changes in soil shortwave absorptance had similar effects, suggesting that some substrata might be less likely to cause high-temperature damage to plants than others. Changes in wind speed at 2 m were more important for the dry soil (where energy dissipation by convection was greater) than for the wet soil, for which evaporative cooling prevailed. However, because the duration of sufficiently wet soil will generally be quite short in deserts, wind speed is likely to be an important variable for the thermal relations of seedlings. Air temperature at 2 m also had an important influence on soil temperatures, a 10 °C change causing an approximately 5 °C change in maximum soil surface temperature (Table 1). Although 10 °C changes in T^{sky} had smaller effects, even a 2 °C or 3 °C difference in maximum temperature may be important to a plant near it thermal tolerance limit.

Maximum soil temperatures were very sensitive to changes in soil water potential between $-0{\cdot}01$ MPa and $-0{\cdot}5$ MPa (Fig. 4), a result of the strong dependency of soil hydraulic conductivity, and hence evaporation, on soil water potential. For example, the hydraulic conductivity decreases 350-fold as ψ^{soil} decreases from $-0{\cdot}01$ MPa to $-0{\cdot}5$ MPa (Young & Nobel 1986). Therefore, although changing the latent heat loss had a substantial influence on the maximum temperatures of the wet soil (Table 1), maintaining high values of latent heat loss depends upon frequent inputs of moisture. The substantial increase in maximum surface temperature with even minor soil drying rapidly reduces the potentially large differences in surface temperatures between dry and wet soils. The increase in maximum temperature as ψ^{soil} decreased below $-0{\cdot}5$ MPa was primarily due to decreases in K^{soil} and secondarily in C_P^{soil} (Fig. 2), rather than to the accompanying decreases in latent heat loss (Table 1). Indeed, J_H^L became a very minor component in the overall soil energy budget when ψ^{soil} was less than $-0{\cdot}5$ MPa. The predicted increase of 3 °C between $-0{\cdot}5$ and $-5{\cdot}0$ MPa can be important to a seedling near the upper limit of its thermal tolerance, indicating that soil moisture could influence seedling survival through effects on temperature as well as the more obvious effects on seedling moisture status.

Seedlings are also influenced by nurse plants, which can modify the local microclimate. For instance, shading by the common desert bunchgrass, *H. rigida*, substantially reduced the simulated soil temperatures (Fig. 5). For every 10% increase in shading, the simulated maximum temperature decreased just over 2 °C at the soil surface and about 1 °C at a depth of 9 cm for both wet and dry soils. For heavily shaded wet soils, the latent heat loss caused the maximum temperature near the soil surface to be lower than for the deeper soil layers. In fact, heavily shaded soils can have maximum surface temperatures below the maximum air temperature at 2 m. The model, which can readily be extended to soils other than the desert soils considered here, predicted the amelioration of high temperatures by nurse plants that can be crucial for the establishment of desert perennials.

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