

Implications of climate change for evaporation from bare soils in a Mediterranean environment

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Abstract The purpose of this study was to predict quantitative changes in evaporation from bare soils in the Mediterranean climate region of Turkey in response to the projections of a regional climate model developed in Japan (hereafter RCM). Daily RCM data for the estimation of reference evapotranspiration (ET_r) and soil evaporation were obtained for the periods of 1994–2003 and 2070–2079. Potential evaporation (E_p) from bare soils was calculated using the Penman–Monteith equation with a surface resistance of zero. Simulation of actual soil evaporation (E_a) was carried out using Aydin model (Aydin et al., Ecological Modelling 182:91–105, 2005) combined with Aydin and Uygur (2006, A model for estimating soil water potential of bare fields. In *Proceedings of the 18th International Soil Meeting (ISM) on Soils Sustaining Life on Earth, Managing Soil and Technology*, Sanliurfa, 477–480pp.) model of predicting soil

water potential at the top surface layer of a bare soil, after performances of Aydin model ($R^2=94.0\%$) and Aydin and Uygur model ($R^2=97.6$) were tested. The latter model is based on the relations among potential soil evaporation, hydraulic diffusivity, and soil wetness, with some simplified assumptions. Input parameters of the model are simple and easily obtainable such as climatic parameters used to compute the potential soil evaporation, average diffusivity for the drying soil, and volumetric water content at field capacity. The combination of Aydin and Aydin and Uygur models appeared to be useful in estimating water potential of soils and E_a from bare soils, with only a few parameters. Unlike ET_r and E_p projected to increase by 92 and 69 mm (equivalent to 8.0 and 7.3% increases) due to the elevated evaporative demand of the atmosphere, respectively, E_a from bare soils is projected to reduce by 50 mm (equivalent to a 16.5% decrease) in response to a decrease in rainfall by 46% in the Mediterranean region of Turkey by the 2070s predicted by RCM, and consequently, to decreased soil wetness in the future.

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Introduction

The globally averaged surface air temperature is projected to increase by 1.4–5.8°C over the period of 1990–2100 (IPCC 2001). Precipitation is not likely

to increase in (semi)-arid regions where the effects of climate change on soil water balance are of major concern as the increased temperature stimulates the evaporative demand of the atmosphere. In such regions, evaporation from the soil surface constitutes a large fraction of the total water loss not only from bare soils but also from cropped fields. Transpiration through crops is regarded as beneficial, but evaporation from bare soils in irrigated fields with a partial canopy cover is considered detrimental. In (semi)-arid regions where summer fallow is practiced, the total amount of water evaporated from the soil may account for most of the incoming precipitation. Evaporation from bare soils has a direct impact on crop yield in rainfed agriculture of (semi)-arid regions. The soil surface remains bare under field crops for many weeks during the periods of seed germination, seedling establishment, and subsequent growth of the young crop when the moisture content of the upper soil layer can be of critical importance. In orchards, the soil surface between the trees is kept bare by frequent tillage and is continuously subjected to evaporation (Mellouli et al. 2000).

Evaporation from a bare soil surface is a complex process including multi-phase transport of soil water to evaporation surface (Alvenas and Jansson 1997; Konukcu et al. 2004). The most important transport processes are characterized by a simultaneous change in the amount of energy or material with time and place (Aydin and Huwe 1993). For example, soil water potential plays a role in water flow, and water potential mainly depends on water content (Campbell 1985). On the other hand, soil evaporation is important as a driving force in water potential changes. Several simplified models have been advocated for the relation between soil evaporation and water potential (Kijne 1974). Ehlers and van der Ploeg (1976), Aydin (1994), and Zhu and Mohanty (2002) estimated evaporation from bare soils as a function of soil water potential (matric potential) at the surface layer, neglecting the influence of the hydraulic gradient. Similarly, Beese et al. (1977), and Huwe and van der Ploeg (1990) quantified the relationship between soil evaporation and water potential at the soil surface.

A simplified model validated by Aydin et al. (2005) (hereafter called Aydin model) to estimate actual soil evaporation using soil water potential measured at the top surface layer appeared to be useful; however, the

objective measurement of soil water potential near the surface of the profile was difficult, especially for drier upper layer. In order to overcome such difficulties, Aydin and Uygur (2006) devised a simple model for predicting soil water potential at the top surface layer. The objective of this study was to predict quantitative changes in evaporation from bare soils in response to climate change in the Mediterranean region of Turkey, based on the combination of Aydin and Aydin and Uygur models and the climate change projections predicted by a regional climate model (RCM) developed in Japan (Sato et al. 2007).

Materials and methods

Description of the models

Daily potential evaporation from bare soils ($E_p = \text{kg m}^{-2} \text{ day}^{-1} \approx \text{mm day}^{-1}$) can be calculated using the Penman–Monteith equation (Allen et al. 1994) with a surface resistance of zero (Brisson et al. 1998; Wallace et al. 1999; Aydin et al. 2005):

$$E_p = \frac{\Delta(R_n - G_s) + 86.4\rho c_p \delta / r_a}{\lambda(\Delta + \gamma)} \quad (1)$$

where Δ is the slope of saturated vapor pressure–temperature curve ($\text{kPa } ^\circ\text{C}^{-1}$), R_n is the net radiation ($\text{MJ m}^{-2} \text{ day}^{-1}$), G_s is the soil heat flux ($\text{MJ m}^{-2} \text{ day}^{-1}$), ρ is the air density (kg m^{-3}), c_p is the specific heat of air ($\text{kJ kg}^{-1} ^\circ\text{C}^{-1} = 1.013$), δ is the vapor pressure deficit of the air (kPa), r_a is the aerodynamic resistance (s m^{-1}), λ is the latent heat of vaporization (MJ kg^{-1}), γ is the psychrometric constant ($\text{kPa } ^\circ\text{C}^{-1}$), and 86.4 is the factor for conversion from kJ s^{-1} to MJ day^{-1} .

Evaporation from soils as a physical process takes place in the aeration zone on the upper surface of the soil (Denisov et al. 2002). Initially evaporation from a wet soil proceeds at the potential rate. With time, the soil surface becomes progressively drier, and drying front moves into the soil. Thus, soil water potential at the top surface layer declines. In order to estimate soil water potential at the top surface layer, a simple model was described by Aydin and Uygur (2006):

$$\Psi = - \left[(1/\alpha) \left(10 \sum E_p \right)^3 / 2(\theta_{fc} - \theta_{ad})(D_{av}t/\pi)^{1/2} \right] \quad (2)$$

where Ψ is soil water potential (cm of water) at the top surface layer, α is a soil specific parameter (cm) related to flow path tortuosity in the soil, ΣE_p is cumulative potential soil evaporation (cm), θ_{fc} and θ_{ad} are volumetric water content ($\text{cm}^3 \text{cm}^{-3}$) at field capacity and air-dryness, respectively, D_{av} is average hydraulic diffusivity ($\text{cm}^2 \text{day}^{-1}$) determined experimentally, t is time (day) and π is 3.1416. The term, $2(\theta_{fc} - \theta_{ad})(D_{av} t/\pi)^{1/2}$, in Eq. 2, represents water supplied from deeper layers to the soil surface. This term is similar to the equation given by Gardner (1959) for the calculation of cumulative actual soil evaporation.

Aydin and Uygur model assumed that (1) water potential near the soil surface decreases as a cubic function of potential soil evaporation based on previous observations in a pot-experiment, and (2) upward transport of water to soil surface dried is capable of partial compensation for such a decrease. During a drying period, the top surface layer of the soil may dry out eventually to air-dry wetness, and the soil surface approaches equilibrium with overlying atmosphere. The soil then no longer evaporates at a considerable rate, except water transport by the slow process of moisture diffusion. The hydraulic diffusivity is a combined term including both water content-matric potential and water content-hydraulic conductivity relationships. The main advantage is that the diffusivity variable is very useful as it accounts for both changes in soil water storage and hydraulic conductivity. The combination of average diffusivity and water content is an estimate of the amount of water that the soil is capable of supplying to the soil surface, assuming an infinite value of potential evaporation rate at the soil surface (Hillel 1980; Hanks 1992).

Assuming that the water potential at dry soil surface is at equilibrium with the atmosphere, the minimum water potential can be derived from the Kelvin equation (Feddes et al. 1978; Brown and Oosterhuis 1992; Kirby and Ringrose-Voase 2000; Aydin et al. 2005):

$$\Psi_{ad} = \frac{R_g T}{mg} \ln H_r \quad (3)$$

where Ψ_{ad} is the water potential for air-dry conditions (cm of water), T is the absolute temperature (K), g is the acceleration due to gravity (981 cm s^{-2}), m is the molecular weight of water ($0.01802 \text{ kg mol}^{-1}$), H_r is the

relative humidity of the air (fraction), and R_g is the universal gas constant ($8.3143 \times 10^4 \text{ kg cm}^2 \text{s}^{-2} \text{mol}^{-1} \text{K}^{-1}$).

Consequently, it is possible to incorporate Eqs. 1, 2, and 3 into Aydin model to compute actual evaporation from a bare soil (Aydin et al. 2005):

$$E_a = \frac{\text{LogI}\Psi I - \text{LogI}\Psi_{ad} I}{\text{LogI}\Psi_{tp} I - \text{LogI}\Psi_{ad} I} E_p \quad (4)$$

where E_a and E_p are actual and potential evaporation rates (mm day^{-1}), respectively, Ψ_{tp} is the absolute values of soil water potential (matric potential) at which actual evaporation starts to drop below potential one, Ψ_{ad} is the absolute values of soil water potential at air-dryness, and Ψ is the absolute value of soil water potential determined by Eq. 2 between the range of Ψ_{tp} and Ψ_{ad} . The values of all Ψ are in cm of water.

Study area

Study site ($36^\circ 59' \text{N}$, $35^\circ 18' \text{E}$) is located in the Cukurova plain (Adana), a southern Mediterranean region of Turkey. The climate of the region is subhumid–semiarid Mediterranean with mild rainy winters and hot dry summers. The long-term mean annual temperature at the site is 18.5°C . The mean annual precipitation is about 700 mm, and the mean relative humidity is 66% (Aydin and Huwe 1993). The soils with different proportions of sand, silt, and clay fractions in the region are predominantly fine-textured soils.

Climate change projections

Climate data for the study location were obtained from DMI (Turkish State Meteorological Service) for the period of 1994–2003 (baseline data). Climate change projections were obtained for the periods of 1994–2003 and 2070–2079 using the RCM developed in Japan (Sato et al. 2007), and daily RCM data were used in the Penman–Monteith equation to calculate reference evapotranspiration (ET_r) and E_p and in Aydin model to obtain E_a .

Pot-experiment

In order to evaluate the performance of the model described by Aydin and Uygur (2006), a pot-experiment

was carried out in a chamber. During the experimental period, air temperature and relative humidity in the chamber were $25 \pm 1^\circ\text{C}$ and $72 \pm 5\%$, respectively. The pots of 182.4 cm^2 surface area with 25.0 cm depth were filled with a clay soil (Sand: 233, Silt: 230, and Clay: 537 g kg^{-1}). Soil was packed in to the columns with average dry bulk density of 1.2 g cm^{-3} .

Two different treatments were applied to pots with two replications. First, the soil in the pots (15.2 cm in diameter and 25.0 cm in depth) without drainage system was kept at saturation throughout the experiment. Thus, potential evaporation was directly calculated from the amount of water loss based on soil surface area. For verification, the daily potential soil evaporation was also estimated from pan evaporation based on a relationship obtained between the saturated surface evaporation and pan evaporation. An evaporation pan (15.2 cm in diameter and 25.0 cm in depth) was used for this purpose. Second, the pots without drainage facility were saturated initially to monitor soil water potential within a drying period. This pot-set was equipped with soil psychrometers. The soil psychrometers were installed at depths of 1 and 5 cm below the soil surface and connected to eight channel-Water Potential System (WESCOR Inc.). For the depth of 1 cm, three replicated soil psychrometers were used. The outputs were continuously recorded using a data logger and converted to MPa. The potentials measured at midday were used.

In this study, α for the experimental soil was taken as 1.1 cm in the calculations of soil water potential by Eq. 2. θ_{fc} was measured as $0.36\text{ (cm}^3\text{ cm}^{-3}\text{)}$, θ_{ad} and D_{av} were assumed to be $0.05\text{ (cm}^3\text{ cm}^{-3}\text{)}$ and $95\text{ (cm}^2\text{ day}^{-1}\text{)}$, respectively, for clay soils. Once the measured and calculated data were fitted, the simulation of soil water potential was carried out continuously for the entire drying period, without additional calibration.

Soil evaporation

For field conditions, ET_r was computed with the Penman–Monteith equation for vegetated surface (Allen et al. 1994; Desborough et al. 1996). Daily potential evaporation from bare soils (E_p) was calculated using the Penman–Monteith equation with a surface resistance of zero (see Eq. 1). In calculations, albedo of bare soil was assumed to be 0.15 (van Dam et al. 1997; Åcs 2003). The data necessary for Eq. 4 are easily obtainable from also the Eqs. 2

and 3. We used 60.0 cm of water as Ψ_{tp} for the clay soil as suggested by Aydin et al. (2005).

For validation, actual soil evaporation was measured under field conditions with three replicated micro-lysimeters installed in the experimental plot (Boast and Robertson 1982; Evett et al. 1995; Yang et al. 2002). The experimental plot with no appreciable slope of 10 m by 10 m was used. The soil is a Vertic Luvisol (Aydin 1994). The micro-lysimeters were composed of open PVC cylinders with 57.0 cm in length, 10.2 cm in internal diameters and a thickness of 0.4 cm . The cylinders, along with a steel cutter, were inserted into wetted soil and then carefully taken out. Wire-gauze sieves were affixed to the bottom of the cylinders in order to retain soil particles without disturbance of drainage. The micro-lysimeters were placed in open-ended PVC columns with 11.4 cm inside diameter in the field, which served as soil-retaining walls. The surface of the soil inside the micro-lysimeters was leveled with the surface of the surrounding soil at the beginning of the experiment (Brisson et al. 1998; Aydin et al. 2005). They were weighed every 2–3 days at the same time with an electronic balance in order to calculate evaporation from changes in weight.

Results and discussion

In order to evaluate the performance for a wide range of soil water potential of Aydin and Uygur model prior to being incorporated into Aydin model, the data from previous experiments were pooled, and the estimated soil water potentials (as suction values) were plotted against measured ones. As can be seen in Fig. 1, the simulated and measured suction values compared reasonably well ($R^2=97.6\%$; $P<0.001$). Similarly, the comparison of measured and simulated soil evaporation was presented based on Aydin model in Fig. 2 ($R^2=94.0\%$; $P<0.001$). Evaporation occurs in two distinct phases in that surface resistance is zero during the first stage, and the evaporation from the soil proceeds at the potential rate. During the falling rate stage, with a dry layer at the surface, the evaporation rate is reduced (Ritchie 1972). The results are comparable with those found by Aydin et al. (2005) for similar soil and climatic conditions. Aydin, and Aydin and Uygur models applied in this study appear to be applicable for a wide range of soils if soil-specific parameters are used.

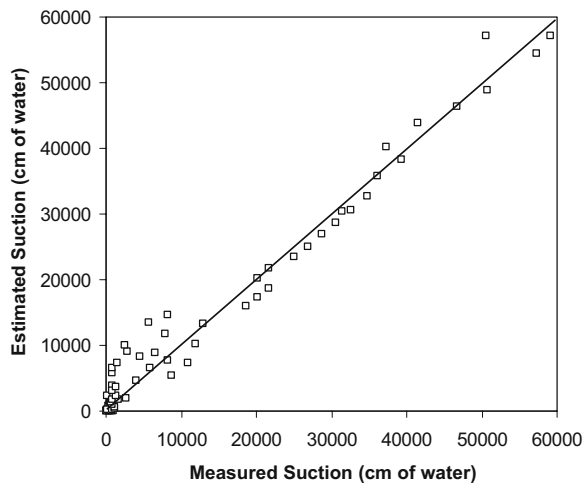


Fig. 1 A comparison of estimated versus measured soil water suction at the depth of 1 cm

Apart from errors in measurements, the systematic errors in the model parameters used should be considered. The main error related to soil water potential measurement is that psychrometer produces a reliable read-out in the range of -0.3 to -6.0 MPa since one single method cannot acceptably cover the full range of soil water (Skierucha 2005). Consequently, Aydin and Uygur model overestimates the soil water suction at the wet range. In a nonlinear system, an average value for diffusivity can be used. This must be a weighted average for many purposes since water movement in the region of highest diffusivity often predominates. Baver et al. (1972)

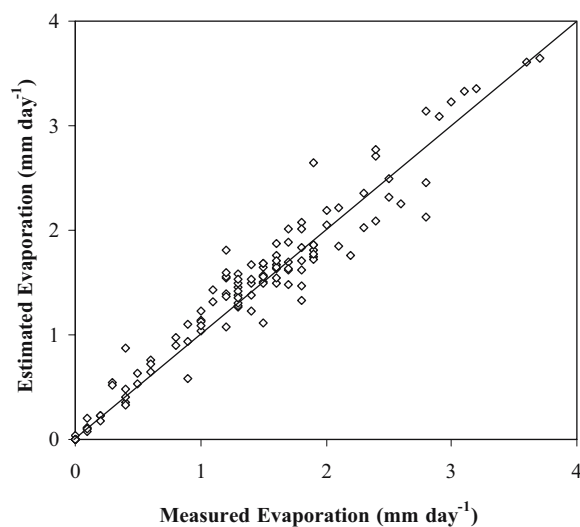


Fig. 2 A comparison of estimated versus measured soil evaporation

reported that a reasonable average diffusivity for a drying clay soil might be $100 \text{ cm}^2 \text{ day}^{-1}$. Thus, a value of $95 \text{ cm}^2 \text{ day}^{-1}$ was used in this study. However, in a drying system, a single and constant “weighted mean diffusivity” is, as such, an oversimplification. For a more realistic treatment, the diffusivity values for different ranges of water content must be determined experimentally.

The other reasons for considerable differences between estimated and measured values could be the complexity of water regime of the clay soil. Similar observations for clay soils were also reported in previous studies (Ritchie et al. 1972; Hasegawa and Sato 1987; Aydin and Huwe 1993; Aydin et al. 2005). In other words, swelling, shrinking and cracking properties of the soils might influence the behaviors of the soil, and consequently the process of water transport in the experimental soil. It is well known that the estimation of water potential of swelling soils is difficult to interpret relative to that of non-swelling soils. In general, estimated and measured soil water suction values were in close agreement, thus indicating that the presented model was valuable in estimating water potential of soils with only a few parameters. However, Aydin and Uygur model has bias for resetting after rainfall events and needs further studies to determine the minimum amount of rainfall based on a ratio of rainfall to potential evaporation to reset the model.

Simulations of daily ET_r , E_p , and E_a rates from bare soils were conducted for each of the years during the periods of 1994–2003 and 2070–2079 and revealed similar patterns. In order to show the relationship among ET_r , E_p , E_a , and rainfall, a comparison of daily changes in the variables for Adana in 1995 is shown as an example in Fig. 3. In general, ET_r rates were higher than E_p values. Kroes et al. (1999) reported that ET_r rates can be multiplied by a coefficient value of 0.5–1.5 to obtain E_p . The evaporation from bare soils depends not only on the atmospheric conditions but also on soil properties. ET_r and E_p rates were lower during the winter months because of the lesser atmospheric evaporative demand. However, E_a rates were mainly found to be a function of the amount and timing of rainfall, and presumably soil wetness in addition to atmospheric evaporative demand.

Monthly variations of ET_r , E_p and E_a for a period of 10 years as of 1994 in Adana are given in Fig. 4. ET_r and E_p were computed using the Penman–Monteith

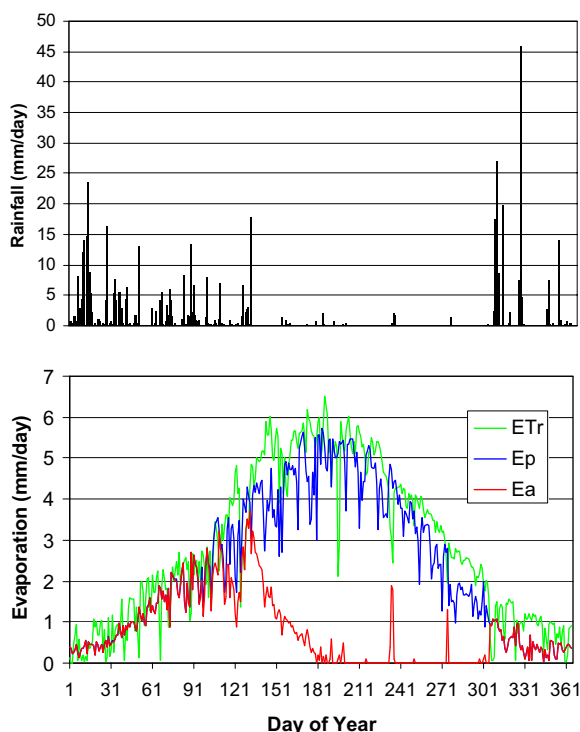


Fig. 3 Comparison of potential (E_p) and actual (E_a) evaporation from a bare clay soil, and reference evapotranspiration (ET_r) along with rainfall in Adana in 1995

equation, and E_a was calculated using Aydin model based on both RCM and observed datasets. Similarly, a comparison of monthly ET_r , E_p and E_a predicted from the RCM data for Adana for the period of 2070–2079 is depicted in Fig. 5. As can be seen in Figs. 4 and 5, the rates of ET_r were overestimated when compared with those of E_p . Both potential rates represent

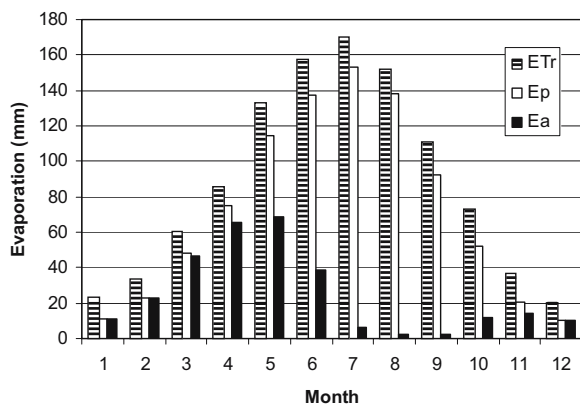


Fig. 4 Monthly mean reference evapotranspiration (ET_r), and potential (E_p) and actual (E_a) soil evaporation calculated from regional climate model (RCM) data for a period of 10 years as of 1994 in Adana

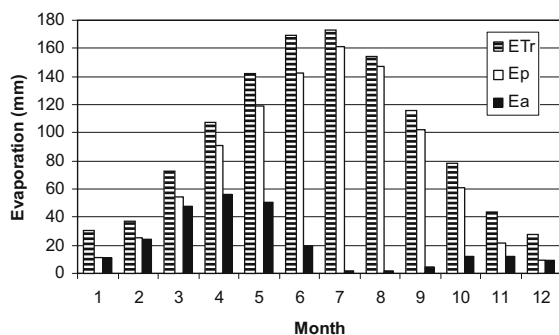


Fig. 5 Monthly mean reference evapotranspiration (ET_r), and potential (E_p) and actual (E_a) soil evaporation calculated from regional climate model (RCM) data for a period of 10 years as of 2070 in Adana

evaporative demand of the atmosphere and were higher during the summer months than the winter months. On the contrary, E_a rates were very low in the summer season and high in the spring months, depending on rainfall pattern and presumably soil wetness.

Changes in the evaporation rates between the present and the future are compared in Fig. 6. Both ET_r and E_p are projected to increase by 92 and 69 mm (equivalent to 8.0 and 7.3% increases), respectively, in response to an increase in temperature by 1.4°C and to changes in other climatic variables such as solar radiation, relative humidity, and wind speed by the 2070s as predicted by RCM (data not shown). Annual E_a is projected to decrease by 50 mm (equivalent to a 16.5% decrease) by the period of 2070–2079 when compared with the baseline period of 1994–2003. Mean annual rainfall is projected to decrease by about 300 mm (equivalent a 46% decrease) by the period of 2070–2079 relative to the baseline period according to RCM. In a warmer climate with lesser precipitation, increased evaporative demand of the atmosphere favors soil dryness.

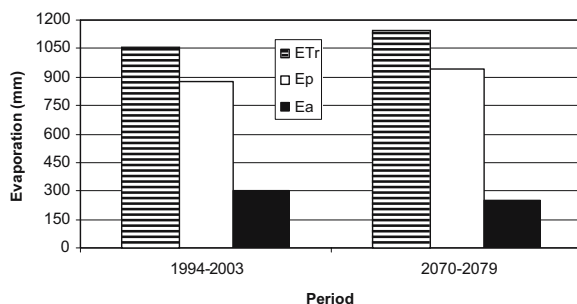


Fig. 6 Changes in annual reference evapotranspiration (ET_r), and potential (E_p) and actual (E_a) soil evaporation between present and future climate in Adana

Simulation of soil evaporation rates from bare soils in a Mediterranean region under a changing climate revealed a reduction in E_a during the hot and dry summer season and an increase in E_a during the mild and rainy spring season, depending on mainly soil wetness. During the wet period of November to May when E_a is close to E_p , the fields should be kept cropped to increase beneficial use of soil water by crops, consequently to prevent water loss through evaporation from soils. Alternatively, adoption of such agronomic practices as retention of crop residues or formation of a natural layer of mulch on the soil surface by proper and timely tillage when evaporation rate is most rapid, therefore, plays a significant role in the magnitude of reduction in soil water loss. Summer fallow is not practiced under the current climatic conditions of the study area but may be adopted in the case of a decrease in rainfall by 46% projected by RCM. In that case, beneficial effects of fallow in terms of the amount of water stored in the soil before planting depend on the timing, length, and type of the fallow period. Fallow with crop residue cover is most likely to be more beneficial in terms of increased soil water storage during the period of highest evaporation rate than that of lowest one.

Conclusions

In the future, potential evapotranspiration from reference crop and potential evaporation from bare soils in the Mediterranean environment was found to increase, whereas actual soil evaporation was found to decrease mainly due to lesser and erratic rainfall pattern, presumably soil wetness. In the study area, rainfall is the source for surface, soil and ground water, among which soil water is the cheapest and most reliable one for crop production when immediate access to surface and ground water is lacking. In a changing global climate, the regions under the Mediterranean climate would increasingly face the challenge of water scarcity and devising innovative and adaptive ecosystem management practices to make efficient use of water available in the soil profile for plant growth. Further studies are needed to assess the full range of climate change impacts and mitigative and adaptation strategies under the projections of regional climate change models.

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