



## Estimating evapotranspiration under warmer climates: Insights from a semi-arid riparian system

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### SUMMARY

This paper presents an approach to quantify evapotranspiration under changing climates, using field observations, theoretical evaporation models and meteorological predictions from global climate models. We analyzed evaporation and meteorological data from three riparian sites located in a semi-arid watershed in southern Arizona USA and found that the surface resistance to water vapor transport was closely related to the vapor pressure deficit. From this, we developed a relatively simple daily conductance model and included a growing season index to accurately replicate the onset and the end of the growing season. After the model was calibrated with observations from January 2003 to December 2007, it was used to predict daily evapotranspiration rates from 2000 to 2100 using Penman–Monteith equation and meteorological projections from the IPCC fourth assessment report climate model runs. Results indicate that atmospheric demand will be greater and lead to increased reference crop evaporation, but evapotranspiration rates at the studied field sites will remain largely unchanged due to stomatal regulation. However, the length of the growing season will increase leading to a greater annual riparian water use. These findings of increased riparian water use and atmospheric demand, likely affecting recharge processes, will lead to greater groundwater deficits and decreased streamflow and have important implications for water management in semi-arid regions.

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### 1. Introduction

The quantification of climate change impacts on hydrology has focused on how changes in precipitation and temperature can affect runoff, evapotranspiration (ET) and recharge (Scibek and Allen, 2006; Seager et al., 2007; Serrat-Capdevila et al., 2007; Milly et al., 2008; Barnett et al., 2008). Nevertheless, in most hydrologic modeling studies attempting to quantify the impacts of climate change, the inclusion of actual ET changes has usually been the least developed aspect of the research. This is mostly due to the complexities of measuring ET, the subsequent lack of data and the number of variables needed to accurately estimate future evaporation rates. The current paper focuses on estimating climate induced changes in the ET of a semi-arid riparian system.

In semi-arid and arid regions evaporation is mostly limited by precipitation in the basin. However, in riparian systems where

there is a linkage between the river and the aquifer, transpiration by riparian ecosystems tapping groundwater is an important component of the water balance in such basins (Goodrich et al., 2000; Scott et al., 2000). Despite this, little effort has been directed to predict changes in evapotranspiration (ET) of riparian ecosystems and vegetation cover. For instance, Serrat-Capdevila et al. (2007) present an approach to link an ensemble of global climate model outputs with a hydrological model. Their work focuses on changes in a semi-arid basin's water budget due to changes in recharge, but they assume yearly ET rates to be constant through the century. Picking on this improbable assumption, the present paper explores the effects of climate change on ET and attempts to fill this void.

Climate change impacts on evapotranspiration can be seen as twofold: (1) changes in ET due to changes in the length of the growing season, and (2) changes in ET rates during the growing season. Most of the modeling studies on climate change impacts in hydrology predict an increase in annual ET due to an earlier start of the growing season, mostly due to earlier snowmelt and a reduction in snow cover (Dankers and Christensen, 2005). Similarly, the end of the growing season, often marked by the first frosts in temperate regions, may be delayed in a warmer climate. Kaszkurewicz

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and Fogg (1967) analyze growing season data for cottonwood and sycamore species from trees widely distributed throughout their natural ranges. Looking at a variety of factors that could potentially influence the beginning and end of the growing season, they study the joint effects of photoperiod and temperature, which would have opposite controls on the growing season. At higher latitudes, the onset of the growing season happens when photoperiods are longer even if temperatures are lower, and vice versa at lower latitudes. Both photoperiod and temperature seem to control growing season onset through a key interplay, allowing for onsets at different temperatures due to different photoperiods. Starr et al. (2000) artificially recreate a longer growing season in a species of forb, using snow cover removal and soil warming. The forbs became active earlier but senesced earlier too. Leaf size and number, photosynthetic assimilation and nutrient concentration remained the same as in the control site. Thus, some plants may not be able to adapt physiologically to an extended growing season and would be at a disadvantage with respect to more adaptive plants or to southern species spreading north due to global warming. However, Churkina et al. (2005) perform a spatial analysis of the relationship between net ecosystem exchange (NEE) and the length of the carbon uptake period (i.e. number of days where the ecosystem is a net carbon sink) and observations seem to show a linear correlation between uptake period and growing season length. Their findings imply that a longer growing season leads to greater carbon uptake and likely, ET. Shi et al. (2008) find tree growth to be strongly affected by temperature at the beginning and end of the growing season, while soil moisture is the main control in between. A percentage of growth limitation for each day of the year is presented as a probability of being growth-limited due to temperature and soil moisture. Menzel and Fabian (1999) analyze more than 30 years of phenological observations in Europe. They report that spring events – such as leaf unfolding – have advanced by 6 days, while fall events – such as leaf coloring – have delayed by 4.8 days, with an average growing season 10.4 days longer than in the early 1960s, attributable to warmer air temperatures. According to their model, more than 70% of the yearly variability in bud-break can be explained by daily temperatures. Similarly, the model predicted an onset advance of up to 6 days per 1 °C increase in winter temperatures.

On the other hand it has proven perhaps even more complex to predict changes in actual evapotranspiration rates during the growing season itself. Jacobs and De Bruin (1997) use a coupled planetary boundary layer and vegetation model to study the effects of doubled CO<sub>2</sub> concentrations on surface resistance and regional transpiration. Their model results show that an increase in surface resistance would be magnified through a positive feedback with the resulting dryer canopy air. Kruijt et al. (2008) provide a good review of the effects of increased CO<sub>2</sub> concentrations on the productivity and functioning of plants. By using partial corrections on crop factors to account for the effects of CO<sub>2</sub> concentrations in stomata conductance and other properties, they estimate future effects on ET in the Netherlands using climate scenarios. Results seem to indicate that reductions of stomatal conductance due to high CO<sub>2</sub> concentrations and higher ET due to warmer temperatures may even each other out. Gedney et al. (2006) present evidence that increasing CO<sub>2</sub> concentrations have in average contributed, through reduced transpiration, to a net increase in runoff. Their findings suggest that reduced stomatal openings have a significant influence in the global water cycle. Using a multi-model approach, Milly et al. (2005) present patterns of trends in streamflow in different continental regions, some are increasing and some are decreasing, as is the case for the semi-arid southwest US. Thus, it is probable that the relevance of findings by Gedney et al. (2006) varies regionally. Wang et al. (2008) quantify changes in Light-Use Efficiency (LUE) and Evaporative Fraction (EF) due to

variations in the ratio of diffuse to total incident solar radiation, which is controlled by cloud and aerosol cover. Because of higher leaf area incidence by diffuse radiation, their findings indicate that LUE can be from 20% to 200% higher with aerosols and increasing cloud cover compared to clear skies. This results in 9% or 15–23% increase in evaporative fraction, the ratio of evapotranspiration to total latent and sensible heat fluxes. Few publications if any have assessed the effects of warmer temperatures and associated meteorological variables on evapotranspiration rates. A physically based approach to estimate future actual evapotranspiration rates is the main contribution of this publication, using insights from field observations, existing evaporation models and climate model projections.

### 1.1. The San Pedro Basin riparian system

The semi-arid San Pedro Basin constitutes one of the last perennial desert rivers in the Southwestern United States. Between its headwaters near Cananea, in Sonora (Mexico) to its confluence with the Gila River 240 km further north in Arizona (US), it hosts the *San Pedro Riparian Natural Conservation Area (SPRNCA)*, a riparian ecosystem and migratory corridor with a high biodiversity. In this semi-arid basin, annual temperature maximums and minimums average 26 °C and 7 °C respectively, and rainfall averages around 350 mm with high spatial and temporal variability. Winter rains from frontal storms provide ~30% of the mean annual precipitation (November to March) and the monsoons – high intensity, short-duration convective storms – along with latter residual moisture from tropical storms, provide ~60% (July to September). From April through June, days are typically very dry and hot. Because of the rainfall regime and the long dry periods between rains, evapotranspiration in the basin is mostly limited by precipitation. In other words, most of the rain that falls in the basin either (1) immediately evaporates back to the atmosphere or runs off as flash floods as is the case for monsoon storms, or (2) is mostly trapped in the soil and evaporates in the following dry period as is the case with winter rains. Only a small part of rainfall in the basin contributes to basin-floor recharge through focused recharge of storm runoff in ephemeral channels, estimated at about 10% or 15% of total basin recharge (Goodrich et al., 2004; Coes and Pool, 2005). However, in basin and range landscapes such as in Southern Arizona, most aquifer recharge originates from rainfall–runoff in the mountains separating the basins, which infiltrates into the sedimentary basin along a fringe at the mountain front. This process is estimated to contribute about 80% of the basin's groundwater recharge (Anderson et al., 1992; Phillips et al., 2004; Wilson and Guan, 2004). The high seasonal and interannual variability of this mountain front recharge is smoothed out by the long travel times of groundwater from the mountain front to the river. Thus, the aquifer, recharged at the mountain front, is able to perennially drain through the river, sustaining a lush riparian ecosystem year-round. Because of this linkage between the aquifer and surface water, riparian vegetation can easily tap ground water along the river, and ET is not limited by precipitation. Riparian transpiration along the river can be an important component of the water balance in such basins (Scott et al., 2008).

Due to the high climatic variability in the region, the vegetation in the area is adapted to cope with strong seasonal changes in the partitioning of surface energy and water fluxes, controlled by water availability, temperature and vapor pressure deficit. While water is thought to be the main limiting factor to evapotranspiration, temperature plays an important role defining the length of the growing season. Measurements in a setting with such a high seasonal variability allowed analysis under a broad range of meteorological conditions and surface controls, as shown in Shuttleworth et al. (2009).

Sustainability problems in the basin appeared with increased groundwater use since the introduction of centrifugal pumps for agriculture in the mid 20th century and rapid population growth in the basin. Lowering groundwater levels could disconnect the aquifer from the river, resulting in its disappearance and that of the protected riparian ecosystem it supports. To address this threat and ensure the water needed for the conservation area a congressional mandate was passed (Public Law 108–136, Section 321) to attain sustainable yield by 2011. The definition of sustainable yield used by the consortium facing its implementation states: “*managing groundwater in a way that can be maintained for an indefinite period of time, without causing unacceptable environmental, economic, or social consequences*”. This implies that in average the sum of annual pumping and riparian consumptive use should be equal or less than annual recharge. Nevertheless, due to increasing temperatures and future climate change impacts affecting the water balance in the basin, sustainable yield may be a moving target.

This paper aims to contribute a quantification of climate change induced variations in both the ET rates during the growing season, and the length of the growing season itself. Experimental eddy covariance and meteorological data from three sites – mesquite woodland, shrubland and sacaton grassland – is used in this study to analyze meteorological controls on riparian evaporation rates in the San Pedro Basin. Our work to quantify ET changes under future climate projections focuses on the evolution of (1) ET rates in our riparian study sites with respective vegetation cover, and of (2) pan evaporation and reference crop evapotranspiration rates. First we analyze data for the period from January 2003 to December 2007, for which the measurements previously described are available (Scott et al., 2004, 2008). Second, in the following section, we design and calibrate a conductance model that captures the behavior of the observed evapotranspiration rates. Third, we estimate future ET rates from 2000 to 2100 by using meteorological projections from a global circulation model and our conductance model.

In the research presented here, we do not address the higher CO<sub>2</sub> concentration effects on stomatal openings. Even if the validity of previous findings remains untested in semi-arid areas (where vegetation is adapted to cope with high atmospheric demands) the research presented here is not antagonistic with them, were these to be verified. While we do not consider these effects or those resulting from potential changes in the ratio of diffuse to total incident sunlight, the findings of this paper are very likely to remain unaffected for semi-arid riparian areas. Since groundwater levels at the three sites were relatively stable throughout the observational period, they were assumed to remain constant. This is not a far-fetched assumption given that river flows (i.e. base-flows, thus groundwater levels) are protected by a law aiming to protect the riparian habitat. Finally, this study also assumes for practical purposes that ecosystem response to climate change will be constant.

## 2. Methods

### 2.1. Study sites and measurements

Experimental data used in this study was gathered by Scott et al. (2004, 2008) and includes measurements of surface energy and water fluxes as well as standard meteorological variables from 2003 to 2007. Measurements were made using eddy covariance towers located at three sites consisting of a mesquite woodland, a sacaton grassland and a mixed mesquite–sacaton shrubland along floodplain terraces of the San Pedro River in southeastern Arizona. The woodland study site is dominated by large velvet

mesquite trees (*Prosopis velutina*). The average tree canopy cover is approximately 75%. The mean canopy height is 7 m, and the depth to groundwater is 10 m. The shrubland site is in a moderately dense stand of velvet mesquite with 55% tree cover. The tree heights are between 3 and 4 m and grasses and various smaller shrubs are abundant in scattered patches in the tree interspaces. The depth to groundwater is 6.5 m. The grassland site is a lush growth of sacaton grass (*Sporobolus wrightii*) and a variety of summer-active herbaceous dicots. The canopy height averages about 1 m with a canopy cover 65% and an average depth to groundwater of 2.5 m. Soil texture profiles at all three sites are similar and consist mainly of gravelly sandy loam layers interspersed with clay and gravel lenses (Scott et al., 2006). In all three sites, vegetation can access groundwater and is only limited by the rate at which it can move water from the roots to the leaves.

### 2.2. Data analysis

For comparison purposes during the data analysis, reference crop rates were calculated with measured meteorological variables at the three sites. Reference crop evapotranspiration (Allen et al., 1998) is the estimated actual ET rate from a well watered, well-specified grass crop, i.e. calculated using well defined stomatal and aerodynamic resistances and using the Penman–Monteith equation (Monteith, 1964).  $\lambda E$  is the ET rate – under prevailing meteorological conditions – of a crop with aerodynamic and surface resistances,  $r_a$  and  $r_s$ , and is expressed as follows:

$$\lambda E = \frac{\Delta(R_n - G) + \frac{\rho_a c_p \text{VPD}}{r_a}}{\Delta + \gamma(1 + \frac{r_s}{r_a})} \quad (1)$$

where  $\Delta$  is the slope of the saturated vapor pressure described as a function of temperature,  $R_n$  and  $G$  are the net radiation and the ground heat flux respectively,  $\rho_a$  is the air density,  $c_p$  is the specific heat of air at constant pressure, VPD is the vapor pressure deficit,  $\gamma$  is the psychrometric “constant” and  $\lambda$  is the latent heat of vaporization. The current paper will stay away from the abstract and often ambiguous term *potential evaporation* and will use only the concept of actual evaporation, as done in Shuttleworth et al. (2009).

For a reference crop the aerodynamic and surface resistances take the following values (Allen et al., 1998):

$$r_a = \frac{208}{u_z}; \quad r_s = 70 \text{ sm}^{-1} \quad (2)$$

Vapor pressure deficit and wind speed measurements were done above the canopy at 14 m in the woodland, 3.5 m in the grassland and mid-canopy at 2.5 m for the shrubland. Thus, equivalent vapor pressure deficits and wind speeds that would be measured at 2 m over a reference crop under the same prevailing meteorological conditions had to be derived in order to calculate reference crop evapotranspiration rates using Penman–Monteith equation. A modified approach from that presented in Shuttleworth (2006) and used in Shuttleworth et al. (2009) was followed, where: (1) equivalent wind speed and vapor pressure deficit are calculated at a blending height of 50 m above the landscape using the aerodynamic properties of the canopy; and (2) the aerodynamic properties of the reference crop are used to calculate the required values back down at 2 m, from those obtained previously at 50 m.

In terms of limiting factors to ET, LAI is the main limitation during the onset of the growing season, while vegetation is leafing-out. Once measured ET reaches a plateau indicating full leaf cover, and before the start of the monsoons, the main limitation to transpiration is the physiologic capacity of the plants to move water up from the water table to their leaf surfaces. During the monsoon season, riparian transpiration is more energy-limited during some

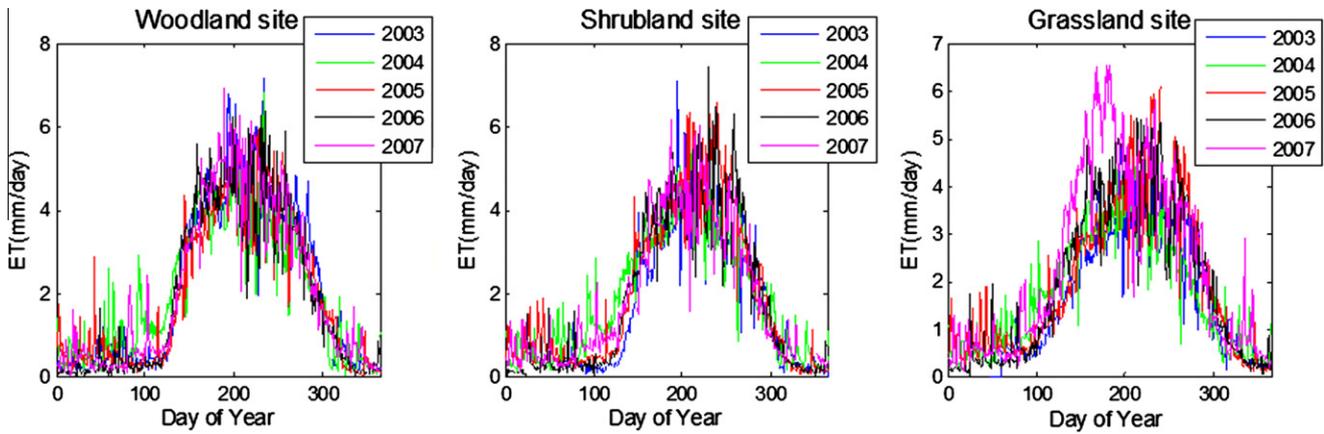


Fig. 1. Evaporation rates in the three sites during the observational period 2003–2007.

days or parts of the day due to cloud cover and monsoon storms. Evaporation can be very important during and after monsoon storms, so low vapor pressure deficits during these hours may also limit transpiration. This can be observed as the peaks and troughs in the ET rate during the summer rainy season (days of year 195–260) in Figs. 1 and 2. Because rainfall events at the study sites in this region are often of high intensity but short duration, we have chosen to deal with evapotranspiration as a whole and consequently use the concept of surface resistance ( $r_s$ ) as a measure of the surface control which includes stomatal resistance as well as soil surface resistance. However, the monsoon season does not start until July. Thus, the rising limb of ET during the first half of the growing season – i.e. the dry season – is dominated by transpiration. When the monsoon starts, some storm related variability can be seen superimposed on the background trend in ET in Figs. 1 and 2. This variability is interpreted as episodic periods of high evaporation following rain events and/or limitations in atmospheric demand on cloudy days.

Yearly evapotranspiration rates (Fig. 1) have been averaged for each site in order to allow for comparison between sites (Fig. 2). Evapotranspiration rates in the three sites are relatively similar, although some differences exist. ET rates during the growing season are higher for the woodland site and lower for the grassland site, with intermediate rates for the shrubland during the monsoon. It is important to note the earlier onset of the growing season

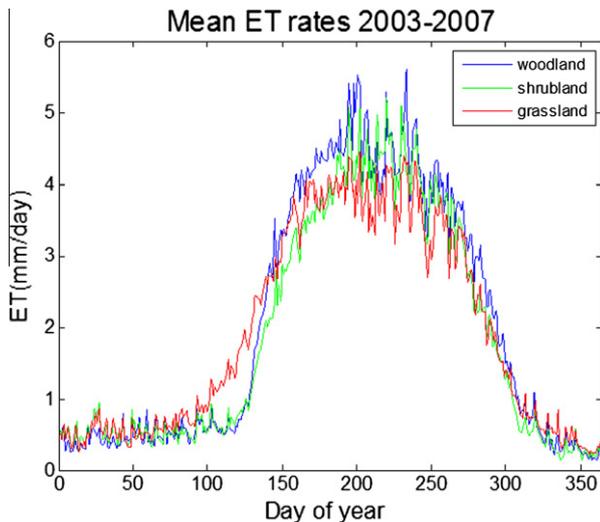


Fig. 2. Average daily evaporation rates for the three sites.

in the grassland, compared to the woodland and shrubland. The end of the growing season happens at the same time for all sites. These differences between sites will have relevant implications for the calibration of our conductance model explained in the following section.

In order to observe the differences between the calculated reference crop estimates and the measured evapotranspiration, daily ET rates were normalized by dividing them by the daily Priestley–Taylor evaporation ( $\lambda E_{PT}$ , (Priestley and Taylor, 1972) and plotted against vapor pressure deficit (VPD) (Fig. 3). In this way, and since Priestley–Taylor equation only depends on available energy, all the reference crop and actual ET rates were normalized by a measure of their available energy. In this way, any controls of VPD on ET can appear regardless of each day's radiation. Priestley–Taylor evaporation is calculated as:

$$\lambda E_{PT} = \alpha \frac{\Delta}{\Delta + \gamma} (R_n - G) \quad (3)$$

The value of  $\alpha = 1.26$  for Priestley–Taylor coefficient was used in the normalization for plotting purposes so that normalized ET rates would tend to unity at low values of VPD, i.e. humid conditions. Fig. 3 shows the normalized ET rates plotted against the vapor pressure deficit at the woodland site.

In Fig. 3 rain days have been removed because of the enhanced rates of evaporation of water intercepted by the vegetation. As monsoon rains are intense and short lived, the movement of adjacent warm dry air in the area due to advective activity can result in higher evapotranspiration rates than it would be physically possible due to radiative inputs. If rain days had been included, a certain number of high values of normalized measured ET would appear at low vapor pressure deficits.

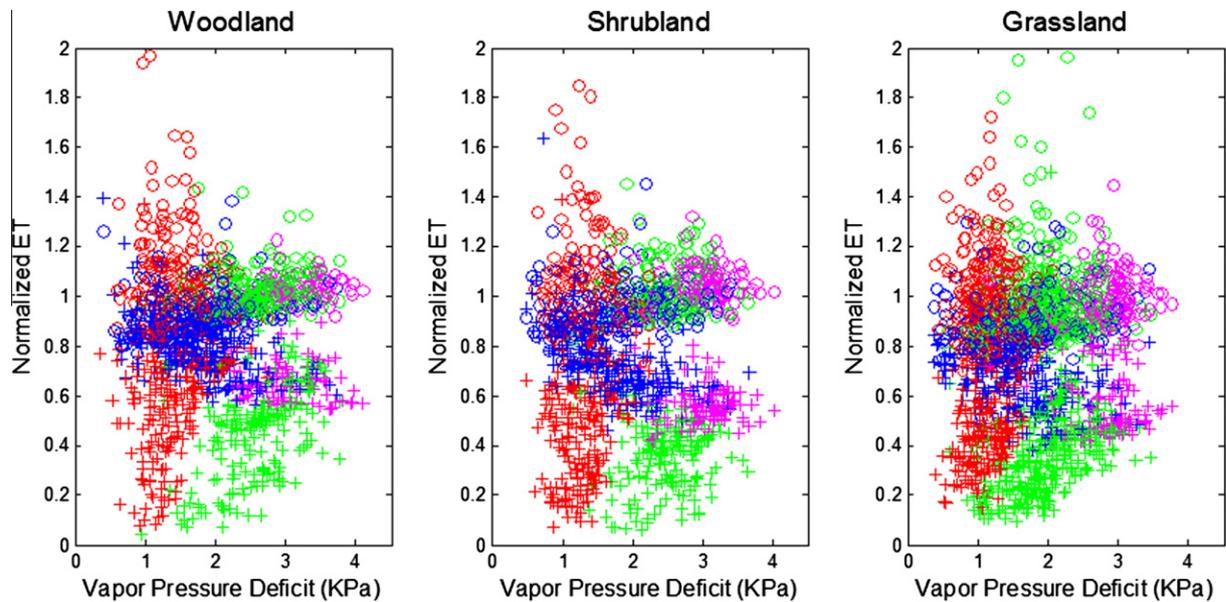
It can be observed that while reference crop evapotranspiration increases with vapor pressure deficit, the measured evapotranspiration rates for the riparian sites decrease. The Penman–Monteith equations for the measured site evapotranspiration ( $\lambda E_m$ ), and for the reference crop evapotranspiration ( $\lambda E_{rc}$ ) can be equated through their common terms ( $\rho \cdot c_p \cdot VPD$ ). Assuming D is roughly the same for both the real and the reference crop, the following expression for ( $\lambda E_{rc}$ )

$$(\lambda E)_{rc} = \frac{\Delta A \cdot [(r_a)_{rc} - (r_a)_c]}{(\Delta + \gamma)(r_a)_{rc} + \gamma(r_s)_{rc}} + (\lambda E)_m \frac{(\Delta + \gamma)(r_a)_c + \gamma(r_s)_c}{(\Delta + \gamma)(r_a)_{rc} + \gamma(r_s)_{rc}} \quad (4)$$

which is an expression of the form:

$$(\lambda E)_{rc} = B + (\lambda E)_m * A(VPD) \quad (5)$$

in which the term  $A(VPD)$  increases with VPD. Since the aerodynamic resistances do not depend on VPD, and the surface resistance



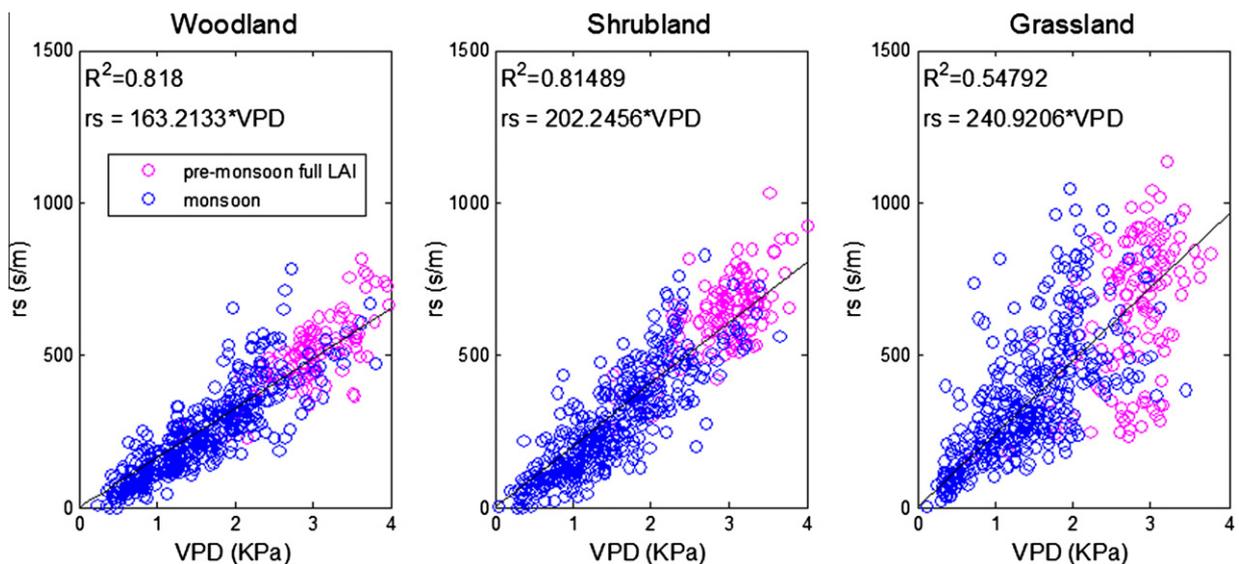
**Fig. 3.** Normalized reference crop (o) and measured site evapotranspiration rates (+) for the three sites. The color code is as follows; green: leaf out period, from the onset of the growing season to full canopy cover; magenta: pre-monsoon period with full canopy cover; blue: monsoon season; red: post-monsoon season until end of senescence. (For interpretation of the references to colour in this figure legend, the reader is referred to the web version of this article.)

of the reference crop is assumed to be fixed at 70 s/m it is easy to see that the term driving the relationship is the surface resistance of the measured riparian crop ( $r_s$ ). Thus, it seems that surface resistance of the real crop increases with VPD, making actual ET decrease.

In order to explore this behavior, we derived surface resistance values from the measured ET rates and calculated aerodynamic resistances of the riparian vegetation using Penman–Monteith equation. Fig. 4 shows the relationship between the obtained daily surface resistance and vapor pressure deficit at the woodland site.

There is a clear linear relationship between surface resistance  $r_s$  and vapor pressure deficit VPD during the mid-stage of the growing season (Fig 4). In Fig. 4, we limited the data to the pre-monsoon and monsoon period to only include times when we could expect that LAI was maximal as this linear relationship is especially clear

when riparian vegetation is not limited by leaf area index. During the leaf out and development stages at the beginning of the growing season and during senescence periods at the end, transpiration is lower and limited by functional leaf area. As seen in Fig. 4, the linear relationship is better for the woodland and shrubland sites and worse for the grassland. This relationship will be used to model daily conductance throughout the year using exclusively vapor pressure deficit data. Correlations of surface resistance and air temperature were not as clear and did not contribute additional information. In fact, for each season, vapor pressure deficit and temperature can be closely related, as shown in Fig. 5, for the three sites. Thus, most of the temperature information is contained in the vapor pressure data, for a specific season of the year. Later on, temperature is used for its implications controlling the onset and end of the growing season, but not evaporation rates within the growing season.



**Fig. 4.** Relationship between surface resistance and vapor pressure deficit for the three sites.

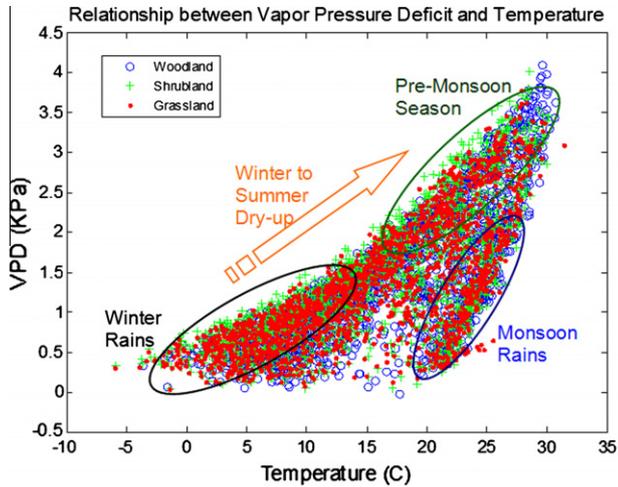


Fig. 5. Relationship between observed vapor pressure deficit (kPa) and temperature (°C) at the three sites and for each season.

### 2.3. Quantifying future ET

Estimates of evaporation rates for the current century, i.e. 2001–2100 have been calculated for a reference crop and for the natural riparian vegetation from our study sites. Meteorological variables predicted for the same period have been obtained from the WCRP CMIP3 multi-model database (<https://esg.llnl.gov:8443/index.jsp>), specifically from “run 4” of the model MPI\_ECHAM5 for Scenario A1B of the IPCC fourth assessment report (IPCC, 2007). Scenario A1B assumes very high GDP growth and energy use, low population growth and land use changes, medium availability of oil and gas, and rapid technological change towards balanced energy sources (fossile and non-fossile). Run 4 was selected due to its high data availability. Daily climate model outputs of temperature, specific humidity, incoming shortwave radiation, incoming longwave radiation, and outgoing longwave radiation were used for this experiment. Although GCM projections of meteorological variables were found to replicate surprisingly well the observational period, a simple downscaling technique was used to ensure that the mean of climate model estimates for the observational period was equivalent to that of the observed data (not shown).

Reference crop ET estimates were calculated using the Penman–Monteith equation, setting surface resistance to  $r_s = 70$  s/m and aerodynamic resistance to that of a reference crop. The aerodynamic resistance between the vegetated surface and a level  $z$  above can be expressed as:

$$r_a = \frac{\ln[(z-d)/z_0] \cdot \ln[(z-d)/(z_0/10)]}{k^2 u_z} \quad (6)$$

where  $u_z$  is the wind speed at height  $z$ ,  $d$  is the zero plane displacement of the vegetated surface,  $z_0$  is its roughness length, and  $k = 0.41$  is the von Karman constant. For a standard reference crop height of  $h_{rc} = 0.12$  m, then  $d = 0.67h_{rc}$  and  $z_0 = 0.123h_{rc}$ . If the equivalent wind speed at  $z = 2$  m is used,  $r_a$  from Eq. (2) applies. Surface wind speed projections were calculated by using the resultant wind speed from the eastward and the northward wind speed components from the climate model.

Temperature projections were used to calculate atmospheric pressure [kPa], as in Allen et al. (1998), air density [ $\text{kg}/\text{m}^3$ ], latent heat of vaporization [ $\text{MJ}/\text{kg}$ ], the psychrometric “constant” [ $\text{kPa}/\text{C}$ ], saturated vapor pressure and its slope versus temperature. Specific humidity and temperature in the GCM projections were used to calculate vapor pressure deficit.

Incoming shortwave ( $S_{in}$ ) and longwave radiation ( $L_{in}$ ) and outgoing longwave radiation ( $L_{out}$ ) were used to calculate the available energy ( $A$ , all in  $\text{W}/\text{m}^2$ ) for each site, assuming an albedo of 0.23 for a reference crop:

$$A_{veg} = S_{in}(1 - \text{albedo}_{rc}) + L_{in} - L_{out} \quad (7)$$

The previous values are used to calculate future reference crop evaporation for the woodland site.

Finally, future evapotranspiration rates for the riparian study sites are also calculated using the previous meteorological projections from the global climate model and a surface resistance model that follows the observed surface resistance for the study period at the three study sites: the woodland, the shrubland and grassland. During a short pre-monsoon period (when mesquite has reached full LAI), the monsoon season and part of the post-monsoon season (before senescence starts) the previous linear relationship between vapor pressure deficit and surface resistance is used:

$$r_s = a * \text{VPD}; \quad \text{thus} \quad g_s = \frac{1}{a * \text{VPD}} \quad (8)$$

where  $g_s = 1/r_s$  is the surface conductance. However, when evapotranspiration is limited by leaf area index or leaf functionality (during leaf-out and senescence) at the beginning and end of the growing season, the previous model consistently overestimates surface conductance, with differences increasing the further away from the peak of the growing season. To account for this, we used a physically sound weighting correction to fit the observed surface conductance year round, by modifying the growing season index proposed by Jolly et al. (2005). Aiming for a global model to assess phenological dynamics, Jolly et al. (2005) combine three indices of environmental limitations (daylength, vapor pressure deficit and minimum temperatures) ranging from 0 to 1, into a single index. Their approach is appropriate for use in climate change conditions as the index quantifies the extent of each climatic limitation within the year and allows the factors to shift and co-limit temporally. For these reasons, it is used here under a modified form, to adjust our conductance model for the entire year as follows:

$$g = \frac{1}{\alpha * \text{VPD}} * \text{iGSI} \quad (9)$$

where  $g$  is conductance (m/s), VPD is vapor pressure deficit (kPa),  $\alpha$  is the slope of the linear relationship between surface resistance  $r_s$  and VPD, and iGSI is our modified growing season index determined as:

$$\text{iGSI} = \text{iTmin} * \text{iPhoto} * \text{iCTemp} \quad (10)$$

Our modified growing season index iGSI is a multiplication of three indices that depend on minimum daily temperature (iTmin), cumulative average temperature since the first day of the year (iCTemp) and photoperiod (iPhoto) calculated as a linear function between zero and one. Each one of these indices depends on two fixed thresholds: below the lower threshold the index equals zero, above the upper threshold, it equals one, and between the two the index goes linearly from 0 to 1. Jolly et al. (2005) propose general values for the thresholds of each index, however, we calibrated them in order to obtain the best fit for our three specific field sites. A total of seven parameters were used in the model: six thresholds (two for each index composing the growing season index) and the slope of the linear relationship between calculated surface resistance and vapor pressure deficit shown in Fig. 4.

It should be pointed out here that the use of crop coefficients and reference crop equations would be misleading to predict future evaporation. It is easy to (1) derive site specific crop coefficients as the ratio of measured and reference crop ET for the period for which measurements are available, and (2) use these crop coefficients to estimate future ET by multiplying them by

reference crop rates under future, warmer climates. However, the use of crop coefficients should be avoided here, as they do not account for changes in surface resistance due to changes in meteorological variables (i.e. vapor pressure deficit), reflecting plant adaptations to climate variability.

In order to calibrate our model and find the optimal parameter set to fit the observations, the Shuffled Complex Evolution (SCE-UA) algorithm was used (Duan et al., 1992). Merging the strengths of the Downhill Simplex procedure (Nelder and Mead, 1965) with the concepts of controlled random search, systematic evolution of points in the direction of global improvement, competitive evolution (Holland, 1975), and complex shuffling, the SCE-UA algorithm represents a synthesis of the best features of several optimization strategies. It is thus a robust global optimization algorithm. Using SCE-UA, the model was calibrated for best fit with all years of the study period. The algorithm searched to minimize an objective function of the mean square error between daily observed and simulated conductance for the observational period 2003–2007.

Fig. 6 shows both the observed and modeled daily surface conductance in the woodland site after calibration for each year of the observation period. As seen, our model does not capture well the peaks in surface conductance that are mainly due to the occurrence of winter rains, as the riparian vegetation is dormant and the evaporation response is not vegetation controlled (e.g., bare soil and interception evaporation). In order to fully capture these peaks our model would need to be more complex. As our main focus here is to assess the impacts of climate change on the growing season's evapotranspiration, we chose to keep the model as it is, capturing well the dynamics of the growing season, rather than making it more parameter-intensive.

Once the model was calibrated and the best parameter set was found for each site, surface resistance values were calculated for the current century. Finally, riparian evapotranspiration from 2000 to 2100 was calculated for each site using the meteorological forcing from the climate model run.

### 3. Results and discussion

During the calibration of our conductance model it was found that minimum temperatures and photoperiod control the end of the growing season. This is in accordance with the fact that during the study period 2003–2007 the growing seasons either end abruptly due to the first strong freeze of the season, or decrease gradually by a certain day in the absence of frost. Thus, as it was also seen during calibration, the end of the growing season can be dictated by either  $iT_{min}$  or  $iPhoto$  each year, but the controls of these two indices appeared to be quite synchronized in average over the long term. As can be seen in Fig. 7 minimum temperatures during leaf out are higher than minimum temperatures during senescence and growing season end.

As leaf-out occurs closer to the middle of the year when photoperiod and temperatures are higher, it is evident that the controls on the leaf out are different from those dictating the end. Cumulative average temperature allows for a very good fit for the onset of the growing season. Given the energy cost required to generate the leaf out, it seems natural that trees wait until later when there is less risk of a hard frost. This is not the case during the senescence period when the leaves can be used until the environment allows. For this reason, it is physically reasonable that most of the increase of growing season length occurs through an earlier onset, which is less limited by photoperiod.

Climate model predictions for the San Pedro basin show increasing temperatures and vapor pressure deficits for the next century (Fig. 8).

Given these predictions, we found the reference crop evaporation will also increase throughout the century (Fig. 9). However, evaporation rates generally do not increase for the riparian sites (Figs. 9 and 10) due to stomatal regulation as a function of vapor pressure deficit. Thus, while the increase in temperature and atmospheric vapor pressure deficit results in increased atmospheric demand (as shown by  $ET_{rc}$ ), surface resistance at the sites increases as

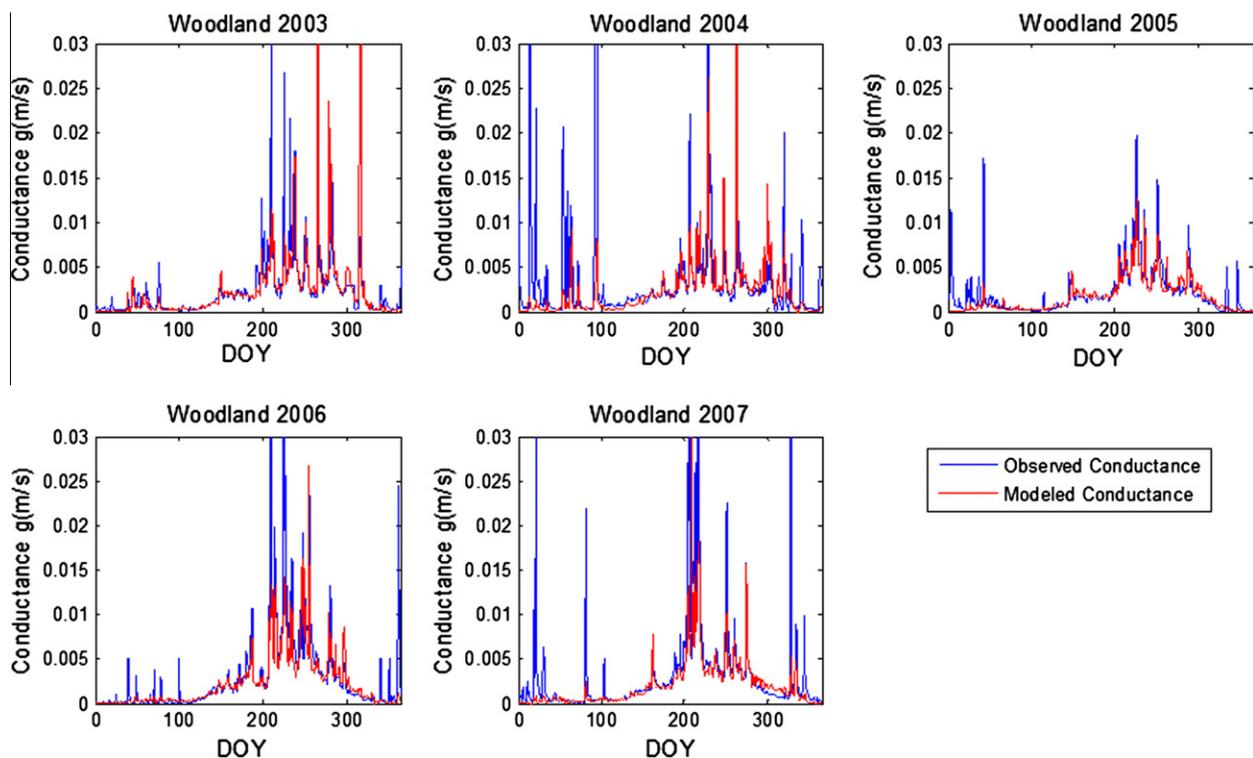


Fig. 6. Observed and modeled surface conductance for the mesquite woodland site (San Pedro Basin).

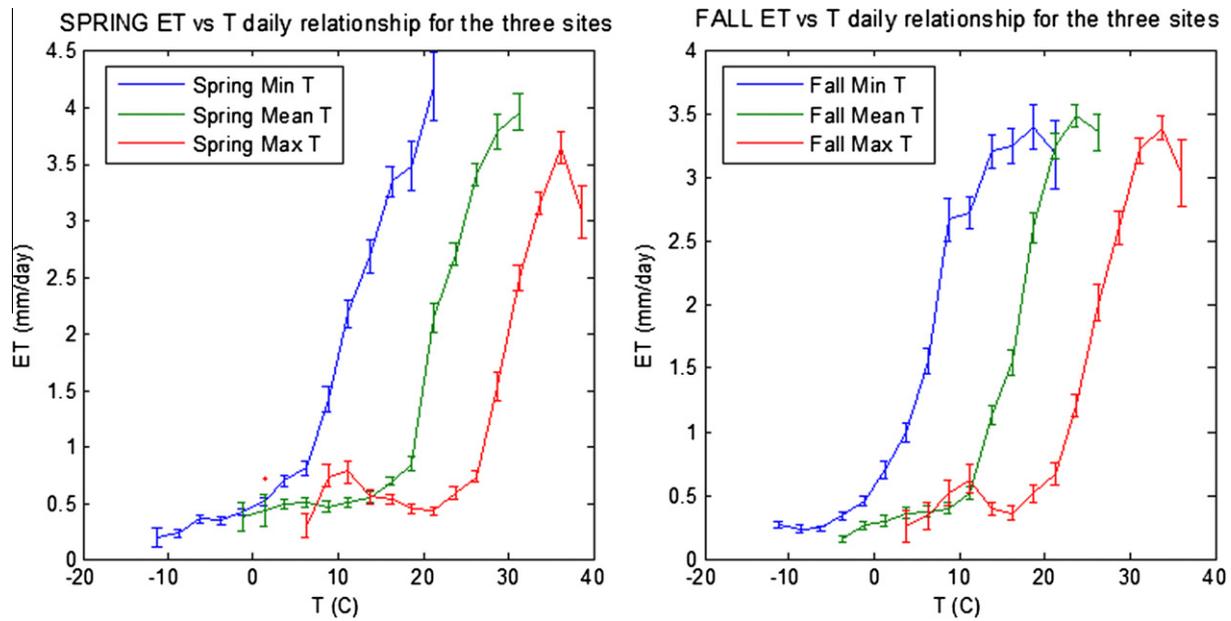


Fig. 7. Relationship between observed evapotranspiration and minimum, average and maximum air temperatures for the onset (spring) and the end (fall) of the growing season for the three sites.

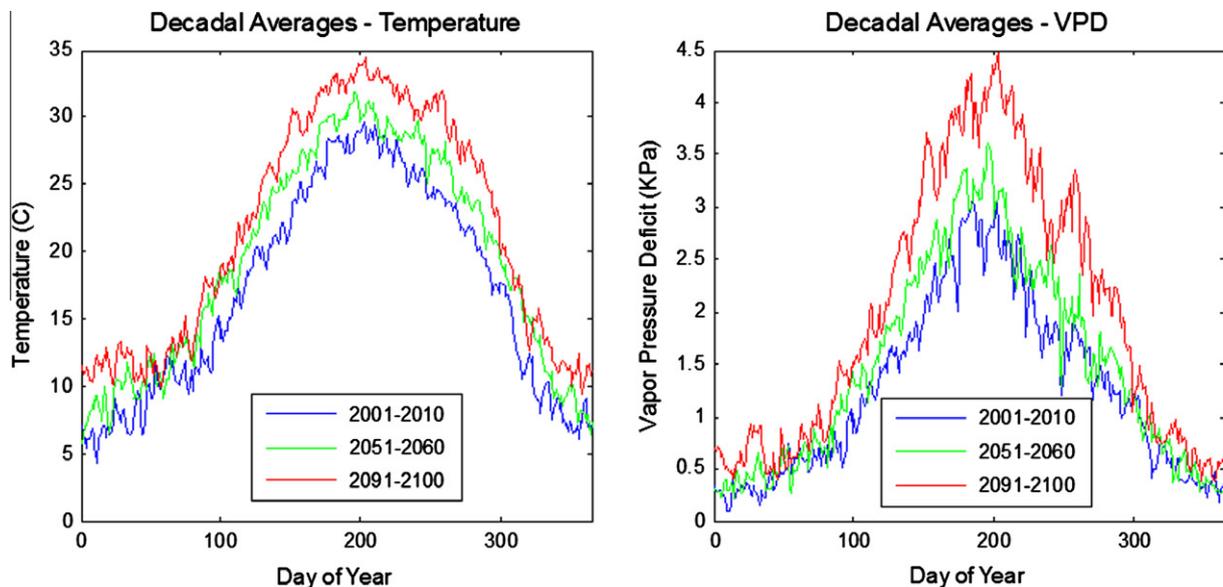


Fig. 8. Decadal averages of temperature and vapor pressure deficit projections for the current century for the San Pedro Basin, from MPI\_ECHAM5, Scenario A1B.

well and results in little change in ET rates. Once the sites reach their full leaf cover, and before the start of the monsoons, the main limitation to transpiration is the physiologic or transport capacity of the plants to pull water up from the water table. Because of this limited physiological capacity, even if water supply is abundant, the trees are “water limited”. This constraint in terms of capacity to pump water up from the water table, combined with very high vapor pressure deficits at the leaf, makes the mesquite prone to xylem cavitation so evidence suggests that mesquite stomata limit water loss during the times of day when temperatures and vapor pressure deficit are high (Scott et al., 2004; Jenerette et al., in preparation). However, even if rates during the growing season will remain the same due to stomatal regulation, the model shows a clear increase in the length of the growing season, notably with a pro-

nounced earlier leaf-out and a somewhat smaller delay to the end of the growing season.

On average the leaf-out is expected to occur 2.6 days earlier each decade for the woodland and the shrubland sites and 2.2 days for the grassland site. The end of the growing season is estimated to be delayed by  $\sim 0.4$  days each decade for all sites. These findings are consistent with the observed lengthening of the growing season in Europe in the last 30 years. Menzel and Fabian (1999) report advances ( $\sim 2$  days/decade) and delays ( $\sim 1.6$  days/decade) in respectively the leaf-out and the senescence of plants that are very similar to the changes predicted in the current research, despite the differences in the study regions and modeling approaches.

Annual ET rates for the next century have been derived for the riparian sites (Fig. 11, for woodland site).

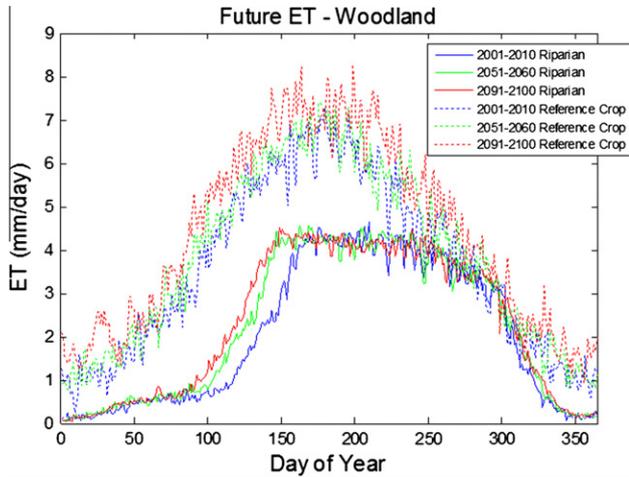


Fig. 9. Reference crop and riparian evapotranspiration averages for the decades 2001–2010 (early), 2051–2060 (mid) and 2091–2100 (late century) at the woodland site.

With the assumption that groundwater levels are stationary, these changes represent an increase of 1.38–1.45% per decade in the total annual water use for the mesquite woodlands and shrublands in the San Pedro basin with similar characteristics to those of the sites, and 1.24% increase for the sacaton grasslands. Such increases in riparian water use are solely due to an earlier onset and a later senescence of the growing season due to predicted warmer temperatures.

Even if the three sites of this study access groundwater and use it to varying degrees, there is also some coupling between vadose zone soil moisture and ET. Yet our model, driven largely by VPD, has been shown to accurately reproduce the observations, implying a high correlation between VPD and surface soil moisture (i.e. dry periods have high VPD and vice versa). Because of the riparian vegetation's access to groundwater, and the strong correlation of VPD with soil moisture and surface resistance, the model is able to accurately reproduce the observations during both dry and wet periods of the growing season, which have varying degrees

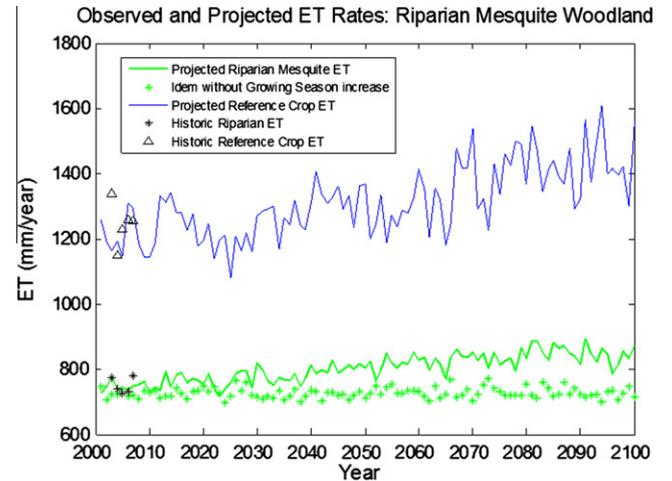


Fig. 11. Calculated evapotranspiration rates for the current century (2000–2100) in the mesquite woodland site, for both the reference crop and riparian ET, accounting for the increase in growing season length (solid line), and without considering the lengthening of the growing season length (dotted).

of evaporation relative to total ET. Hence, temperature and VPD projections alone are able to provide, in the context of the current study, reasonable estimates of future ET.

Looking at ET partitioning and carbon fluxes in the same mesquite woodland site as the one used for this study, Yopez et al. (2007) report a T/ET ratio of 1 for the dry season before the arrival of the monsoon rains, i.e. all ET is derived from aquifer water. Overall in their study year 2002, the mesquites satisfied their water needs with 80% of groundwater and 20% by monsoon precipitation. With negligible runoff and no recharge at the site, the rest of the rainfall was divided equally between evaporation and transpiration by the understory vegetation. Thus, our projected increase in riparian ET due to a lengthening of the growing season will come in its entirety from groundwater. The basin aquifer outflows of pumping, ET and streamflow already exceed recharge in the basin; thus, additional riparian water use will further tax the long-term ecological viability of the riparian habitat.

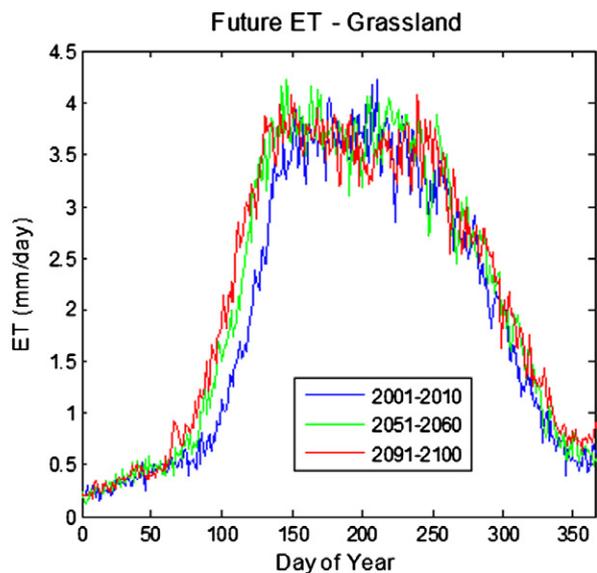
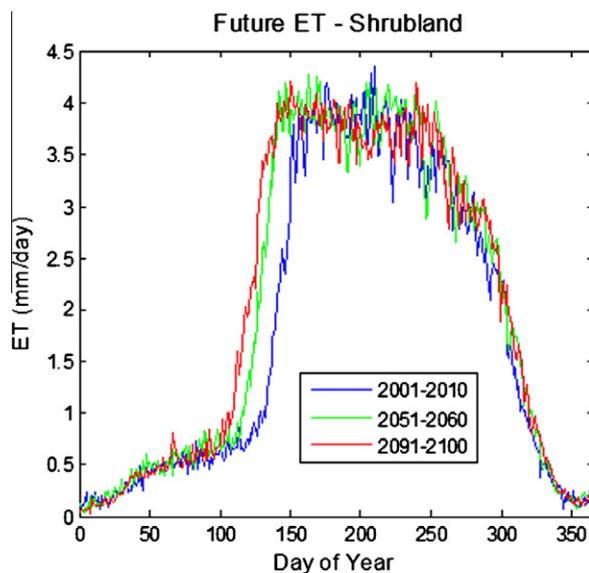


Fig. 10. Riparian evapotranspiration averages for the decades 2001–2010 (early), 2051–2060 (mid) and 2091–2100 (late century) at the shrubland and grassland sites.

#### 4. Conclusions

The current paper presents an approach to quantify future evapotranspiration rates accounting for changes in the length of the growing season and the control of vegetation on evaporation rates. Eddy-covariance and meteorological field observations in three riparian sites where vegetation is able to access groundwater in the San Pedro basin have been used to develop and calibrate a simple conductance model. The conductance model is based on a linear relationship between surface resistance and vapor pressure deficit and a growing season index modified from Jolly et al. (2005), ranging from 0 to 1. The model could replicate accurately the conductance of the three sites for the study period by mimicking the growing season. It was found that minimum temperature and photoperiod controlled the end of the growing season, while cumulative temperatures captured the leaf-out. Using meteorological estimates from the IPCC fourth assessment report (MPI\_EC-HAM5 model run 4 under Scenario A1B) daily evapotranspiration rates were calculated for the current century. Minimum and average temperatures, and vapor pressure deficit projections were used to drive the conductance model. The simulated conductance was then used in the Penman–Monteith equation, driven also by other meteorological estimates for the current century, and future ET rates were obtained.

It was found that the reference crop evapotranspiration will increase with predicted higher atmospheric demand (i.e. greater vapor pressure deficits and higher temperatures). However, the model predicts that ET rates during the growing season at the riparian sites will remain the same due to stomatal regulation. This is reflected in the model by the observed relationship between surface resistance and vapor pressure deficit and is coherent with the fact that native vegetation is well adapted to regulate water loss in semi-arid environments with atmospheric extremes. Another important finding is that the length of the growing season will gradually increase due to the increase in temperatures. Most of the increase will be due to an earlier leaf-out at the onset, and to a lesser extent by a small delay in the senescence. This result is logical since the end of the growing season is also controlled by the photoperiod, which remains constant through the century, while the onset of the growing season currently happens at higher photoperiods and is controlled only by cumulative temperature. In average the leaf-out is expected to occur in average 2.6 days earlier each decade for the woodland and the shrubland sites and 2.2 days for the grassland site. The end of the growing season is estimated to be delayed by 0.44 days each decade for all sites, in average. With the assumption that groundwater levels are stationary, these changes represent an increase of 1.38–1.45% per decade in the total annual evapotranspiration rates for the mesquite woodlands and shrublands in the San Pedro basin with similar characteristics to those of the sites, and 1.24% increase for the sacaton grasslands. These findings are important in regards to groundwater use, since all transpiration from the onset of the growing season until the arrival of the monsoons is entirely fed by groundwater. Thus, the previous increases will reflect directly on the groundwater budget of the basin (Scott et al., 2008). Increases in estimates of reference crop evaporation obeying to higher atmospheric demand have important implications with regards to groundwater recharge. Adding to its predicted decrease due to lesser rainfall (Serrat-Capdevila et al., 2007), higher evaporation potentials in the basin are likely to further reduce recharge rates.

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