Hydrology and Earth System Sciences Discussions

Calculation of reference evapotranspiration surfaces in distributed hydrological modelling at different temporal scales

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Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

◄◄ | ►►

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

Abstract

Each individual process in the soil water balance affected by evaporation processes has a certain representative temporal scale (e.g. canopy evapotranspiration, snowmelt or soil water loss). In this study, the implementation in distributed hydrological modelling at cell scale of the ASCE-Peman-Monteith (ASCE-PM) equation is proposed at hourly time steps whilst the Hargreaves formulation was considered to be appropriate for the water and energy balance at a daily scale due to its simplicity of application once the distributed values of temperature are available at cell scale. However, the coefficient of Hargreaves equation must be previously calibrated. The interplay of different factors at different temporal scales became evident in the calibration process at the local scale of weather stations. However, the best fits against daily estimates by ASCE-PM were achieved when differentiating between the wet and the dry season. For the spatial distribution of Hargreaves coefficient at watershed scale, a regionalization in the area around each weather station was proposed in terms of areas of influence. The best results at watershed scale were obtained after a spatial correction for alpine areas, when the average of the difference cell by cell between ASCE-PM and Hargreaves distributed daily estimates were 0.02 and 0.15 mm day^{-1} for the wet and the dry seasons respectively. In all the cases, the best interpolation results were obtained using $C - I$ (calculate and interpolate) procedures.

1 Introduction

Evapotranspiration, as a component of the soil water balance, plays an important role in the environment at global, regional and local scales. The calculation of evapotranspiration is of major concern for regional management and scheduling of irrigation, reservoir operation studies, capacity of channel design, studies of agricultural potential, effects of land use, changes in water bodies, etc. (Hargreaves and Allen, 2003; Irmak et al., 2003; Maeda el at., 2010). Besides, evaporation processes are extremely

important in distributed hydrological modelling and often constitute the dominant hydrological process in terms of total amount in the water-mass balance (Beven, 1979), especially in arid and semiarid zones (Vicente-Serrano et al., 2007). The response of the watershed to an individual rain storm is scarcely affected by evaporation processes during the storm, and, therefore, their influence is often disregarded. However, during periods between storm events such losses are extremely important as they determine the antecedent soil moisture content, prior to the next event and, therefore, the runoff generation capacity (Gómez-Plaza et al., 2001). Thus, hydrological models that simulate continuously beyond the event scale should carefully consider these water losses as a component of the water balance in the interstorm periods (Maneta et al., 2008).

The application of the different formulations of the Penman-Monteith combination equation (PM) for the computation of evapotranspiration over a reference surface, ET_0, has been extensively used all over the world (Allen et al., 1998; Garcia et al., 2004). The PM equation presents two main advantages over the rest: (1) it is physically-based and, therefore, can be globally applied without any need to estimate additional parameters; (2) it is well documented, implemented in a wide range of software, and has been calibrated by means of varied lysimeters (Droogers and Allen, 2002). This is the reason why it is frequently cited as the preferred method for the calculation of ET_0, especially for calculations at short temporal scales (Alexandris and Kerkides, 2003). Thus, the good results obtained in many different studies at daily and even higher scales is surprising considering that the combination equation was theoretically derived for instantaneous values of the variables involved (Allen et al., 2006). However, the main drawback of this method is its great demand of input data: air temperature, wind speed, relative humidity, and solar radiation. The number of weather stations where all these variables are recorded hourly, or even daily, is limited in many areas of the globe (Irmak et al., 2003; Gavilán et al., 2006). Despite the attempts of Allen et al. (1998) to estimate solar radiation and humidity from other variables easier to measure, it is difficult to obtain the required accuracy without modern electronic devices, especially those concerning wind speed and air vapour pressure values. Besides, the lack of

Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

◄ | ►

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

reliable measurements in areas where ET_0 estimates are especially needed is very common (Allen and Pruitt, 1986; Liu and Todini, 2002; Hargreaves and Allen, 2003; Maeda et al., 2010). These shortcomings in the application of the combination equation motivated the derivation of less demanding models in terms of input data such as the Hargreaves equation, where just daily maximum and minimum temperature values are required (Hargreaves et al., 1985); Jensen-Haise's method, which estimates ET_0 as a function of the daily mean temperature and mean incident solar radiation (Irmak et al., 2003); or the Blaney-Criddle formula that only requires temperature and day length data (Allen and Pruitt, 1986).

Hargreaves method can be considered as a semi-empirical approximation as it incorporates extraterrestrial radiation in combination with temperature as indicators of global radiation, and the daily temperature range as an indicator of humidity and cloudiness (Shuttelworth, 1993; Stefano and Ferro, 1997). Cloudiness is inversely related to the temperature range, and the influence of relative humidity is also related to that range, as there is a linear relationship between both variables (Hargreaves and Samani, 1982). According to Hargreaves, the incorporation of the temperature range in the equation compensates for the influence of advection as it depends on the interaction of temperature, relative humidity, vapour pressure and wind speed, all of them related to the temperature range (Hargreaves and Allen, 2003). The calibration of this method with data from high quality lysimeters under a wide range of climatological conditions demonstrated that the accuracy of the method was similar to PM for ET_0 estimations at weekly and even longer time steps (Hargreaves, 1994; Droogers and Allen, 2002; Vaderlinden et al., 2004). However, accurate ET_0 estimations at daily scale have been reported in literature (Hargreaves and Allen, 2003). Thus, its use has been recommended when there is a lack of reliable data (Droogers and Allen, 2002; Jabloun and Sahli, 2008). Besides, temperature can be reasonably interpolated in areas where measurements are scarce. Jensen et al. (1997) recommended Hargreaves equation as the easiest and most accurate empirical method at stations where standard reference conditions are not present, when not all the variables required in PM

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

are measured or in situations where measurements have errors, especially concerning relative humidity data.

Nevertheless, the application of Hargreaves formula in arid zones must be done with caution as the thermal range does not completely consider the aerodynamic terms (García et al., 2004). In general, reference methods such as Hargreaves, Blaney-Criddle, etc., present some degree of empiricism and do not include all the environmental processes involved in the evapotranspiration process. Therefore, on the one hand, calibration and validation must be done at the local scale (Allen and Pruitt, 1986; Droogers and Allen, 2002; Maeda et al., 2010). On the other hand, within the scope of distributed hydrological modelling, evapotranspiration surfaces as inputs to the models need to be generated. However, in mountainous areas where the monitoring network ineffectively covers the complex heterogeneity of the terrain, simple geostatistical methods do not always provide a representative enough spatial interpolation of evapotranspiration estimates at weather stations, and a detailed study of the features that create strong local gradients must be applied (Herrero et al., 2007; McVicar et al., 2007; Vicente-Serrano et al., 2007; Aguilar et al., 2010). In this context, the application of the Hargreaves equation requires an analysis of the spatial distribution if its parameter (c_H).

Finally, since the implementation of evapotranspiration calculation in continuous hydrological modelling must include the different temporal scales representative of each individual process involved (e.g. canopy evapotranspiration and snowmelt at hourly scale, soil water loss immediately after a rain storm at hourly scale, soil water loss in very dry periods at daily scale, etc.), further analyses should be carried out to include such scale effects.

The purpose of this study is to develop a spatio-temporal procedure for the calibration of Hargreaves equation in heterogeneous watersheds oriented towards the distributed hydrological modelling. Thus, calibration by aggregation of ET_0 at different time intervals is applied in order to consider the influence of wet and dry periods within the year commonly found in Mediterranean areas. Then, different spatial interpolation

Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

|◄ | ►|

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

procedures are tested according to the spatial distribution of the weather stations available in the study area.

As the PM equation can be considered as being the most physically realistic model for the computation of evapotranspiration (Liu and Todini, 2002), it is proposed for the water and energy distributed balance at any time scale once input datasets to the equation are available at cell scale. However, the Hargreaves formulation is proposed at a daily time scale due to its simplicity of application once distributed values of temperature have been generated and the Hargreaves coefficient (set to 0.0023 in the original equation), c_{H} in the foregoing, has been locally calibrated (Jabloun and Sahli, 2008).

2   Materials and methods

2.1   Study area and data sources

The study area is the Guadalfeo river watershed, Southern Spain (Fig. 1), where the highest altitudes in Spain can be found (3482 m) with the coastline only 40 km away, in a 1300 km^2 area, which results in the interaction between semiarid Mediterranean and alpine climate conditions, with the regular presence of snow. The combination of these altitudinal gradients together with the large number of vegetation, landforms and soil types produces a complex mountainous terrain with a variable hydrological behaviour. The main part of the watershed, in terms of hydrology, is comprised of the southern hillside of Sierra Nevada, where global radiation is high throughout the year due to its southern orientation and lack of cloud cover, even during winter (Aguilar et al., 2010). Thus, a considerably high evaporative demand is present in this part of the watershed throughout the year (mean annual values of ET$_0$ close to 1000 mm year^{-1}). In addition, the presence of snow in the Northern area constitutes both a delayed water supply to the system as snowmelt, and an evaporative source under certain conditions (Herrero et al., 2009; Millares et al., 2009).

The topographic input data are represented by a digital elevation model (DEM) with a horizontal resolution of 30 x 30 m and 1 m of vertical precision (Fig. 1).

The meteorological data used in this study consisted of hourly datasets from 11 November 2004 to 2 July 2010 of the variables involved in the equations provided by the three automatic stations of the Agroclimatic Information Network of Andalusia (RIA) available in the watershed (referred to as 601, 602 and 603 in Fig. 1), as well as hourly datasets recorded at a new weather station available from 2004 in Sierra Nevada by the University of Granada Environmental flow dynamics Research Group at an elevation of 2510 m (802 in Fig. 1). Besides, daily datasets from three stations of the Andalusian Alert and Phytosanitary Information Network (RAIF) were available in the area (701, 702 and 703 in Fig. 1). Specific algorithms to interpolate the meteorological input data summarized in Herrero et al. (2007) were applied in order to avoid the limitations that standard methods pose in abrupt topography. Thus, topographic corrections were used for temperature, precipitation and solar radiation (Susong et al., 1999; Garen and Marks, 2005; Herrero, 2007; Aguilar et al., 2010).

The calibration process was carried out with datasets from 11 November 2004 to 31 August 2008. Then, validation was applied with datasets from 1 September 2008 to 2 July 2010.

2.2 Generation of distributed surfaces of evapotranspiration

To derive daily distributed surfaces of ET_0 which were consistent with the complex topography in the watershed, the following steps were followed: (1) PM equation was used to calculate ET_0 at stations 601, 602, 603 and 802, where detailed hourly meteorological datasets were available. (2) These results were used as reference values for the local calibration of Hargreaves equation at each station. (3) Then, a temporal calibration of Hargreaves equation was achieved in order to consider the influence of wet and dry periods on the goodness of fit of the results. (4) Finally, an analysis of the best procedure to spatially interpolate Hargreaves equation is provided.

2.2.1 Penman-Monteith equation

The Penman-Monteith (PM) combination equation has a strong theoretical basis that combines in the first term an energy balance accounting for radiation and soil heat transfer, and in the second term a function for the aerodynamic transport that quantifies the vapour advective displacement of the evaporative surface (Allen, 1986; Stefano and Ferro, 1997). After the sensitivity analysis had demonstrated a strong influence of the aerodynamic and canopy resistance coefficients in the estimates of ET (Beven, 1979), a reference surface was defined (Shuttleworth, 1993; Pereira et al., 1999) in order to "standardize" the initial conditions and neglect the influence of crops and soil characteristics in the estimations of evapotranspiration, leading to the FAO PM equation for the calculation of the daily reference evapotranspiration, ET_0. Then, the parameterization for hourly time-steps calculations led to the equation known as FAO56-PM (Allen et al., 1998). Finally, in order to unify criteria concerning the reference surface and to simplify and clarify the application of the FAO56-PM equation, the ASCE derived the ASCE-PM equation (Eq. 1) including the variation of the resistance coefficients depending on the reference crop, the temporal time-step and, for hourly time-steps, different values for daytime and night time. This modification to the FAO56-PM allows the evaluation of the effects due to changes in wind speed, air temperature and vapour pressure deficit throughout the day (e.g. Gavilán et al., 2008). ASCE-PM equation can be expressed as (e.g. Itenfisu et al., 2003; Gavilán et al., 2007):

$$ET_0^{ASCE} = \frac{0.408\Delta(R_n - G) + \gamma\frac{C_n}{T+273.16}u_2(e_s - e_a)}{\Delta + \gamma(1 + C_d u_2)} \tag{1}$$

where ET_0^{ASCE} is the reference evapotranspiration during each time step (mm Δt^{-1}); Δ the slope of the vapour pressure-temperature-curve saturation calculated at mean air temperature (kPa $°C^{-1}$); γ the psychrometric constant (kPa $°C^{-1}$); R_n and G, the net radiation and soil heat fluxes, respectively, both in mm Δt^{-1} water equivalent; e_a and e_s the actual and saturation vapour pressure (kPa), respectively; T the daily mean

Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

air temperature (°C) and u_2 the wind speed, both measured at a height of 2 m above the soil surface (m s^{-1}). Finally, C_d and C_n are resistance coefficients, which vary with the reference crop, temporal time-step and, in the case of hourly time-steps, with daytime and night time. Table 1 shows their values at different time steps for the same reference crop as in the FAO56-PM. The calculation of some of the variables involved in the ASCE-PM equation can be found in detail depending on the available input data in Allen et al. (1998).

The available energy at the soil surface is the first control of the process, so the estimation of this flux from available data sometimes conditions the selection of a method to generate evapotranspiration values (Shuttleworth, 1993). In the present study site, a topographic model for the estimation of global radiation at the cell scale was applied (Aguilar et al., 2010), whereas net long wave radiation was computed from the atmospheric emissivity, which was calculated through a parametric expression by Herrero et al. (2009) based on near-surface measurements of solar radiation and relative humidity, valid for the local conditions of the study area.

2.2.2 Hargreaves equation

Hargreaves et al. (1985) developed their empirical alternative approach from the combination of Eqs. (2) and (3).

$$\text{ET}_0^{\text{HG}} = 0.0135 \cdot R_g(T + 17.8) \quad \text{(Hargreaves et al., 1985)} \tag{2}$$

$$R_g = c \cdot R_o \sqrt{\Delta T} \quad \text{(Hargreaves and Samani, 1982)} \tag{3}$$

where T is the daily mean temperature (°C), R_g the global solar radiation in mm day^{-1} evaporation equivalent, c an empirical coefficient that after calibration was fixed to 0.16, R_o the extraterrestrial solar radiation in mm day^{-1} evaporation equivalent, and ΔT the daily temperature range ($\Delta T = T_{\text{max}} - T_{\text{min}}$, T_{max}, T_{min} the daily maximum and minimum

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

HESSD

8, 4813–4850, 2011

Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

◄ | ►

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

temperature values, respectively). Finally, the combination of Eqs. (2) and (3) yields the Hargreaves equation (Hargreaves et al., 1985) for ET_0 (mm day^{-1}):

$$ET_0^{HG} = c_H \cdot 0.408 R_o \cdot (T + 17.8) \sqrt{T_{max} - T_{min}} \qquad (4)$$

where the value of 0.408 is the inverse of the latent heat flux of vaporization at 20 °C,
5 changing the extraterrestrial radiation units from MJ m^{-2} day^{-1} into mm day^{-1} of evaporation equivalent (Allen et al., 1998), and c_H is the Hargreaves coefficient.

Hargreaves et al. (1985) stated that Eq. (2) tends to overestimate evapotranspiration in coastal areas where the daily mean temperature is relatively high, whereas the coefficient c in Eq. (3) must be increased in such cases as the daily temperature range
10 is lower due to the "attenuation" of the sea, and decreases at high elevation locations, particularly in mountainous valleys, where air mass movements down slope increase the temperature range. Therefore, as the errors in Eqs. (2) and (3) are of the same order of magnitude but opposite in sign, the value of c_H in Eq. (4) can be fixed to 0.0023 when a standard c value without local calibration is considered compensating for the
15 errors in both equations (Hargreaves et al., 1985; Hargreaves, 1994).

2.2.3 Temporal correction analysis

Firstly, hourly ET_0 values were calculated at weather stations with hourly available datasets in the study area through ASCE-PM equation (Eq. 1) for the period between November 2004 and August 2008, in order to include the alpine sites conditions given
20 by the installation of station 802 in November 2004. Then, hourly values were aggregated to obtain daily values (ET_0^{ASCE}) and compared with daily ET_0 estimates calculated by ASCE-PM equation, similar to FAO56-PM at daily scale (ET_0^{FAO}), but assuming the particularities of the daily computation as stated before (negligible G values, and constant resistance coefficient). This comparison reflects the method's internal consis-
25 tency and robustness when applied at different temporal scales (Itenfisu et al., 2003; Gavilán et al., 2008). The results were evaluated by means of the Mean Error, ME

Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

(mm day^{-1}), and the Root Mean Squared Error, RMSE (mm day^{-1}), as indicators of the deviation and accuracy in the estimations at each station as suggested by Jacovides and Kontoyiannis (1995) for the evaluation of ET_0 estimations. Also, the mean absolute error, MAE (mm day^{-1}), and the ratio between mean estimations, R, were computed.

Once the consistency of the ASCE-PM method was analysed, and considering the error values obtained from this comparison, the daily time step in the application of PM equation was selected (ET_0^{FAO}) and thus, the calibration process was applied in the 7 weather stations with daily datasets available in the area (Fig. 1). Therefore, local calibration of c_H was performed at each station against the daily estimates by ASCE-PM approximation at different aggregated temporal scales, and error values were compared in order to evaluate the different alternatives.

The use of PM ET_0 estimates as a pattern for the calibration of the Hargreaves equation has been previously done by other authors from different versions of Penman-Monteith equations (Allen and Pruitt, 1986; Hargreaves, 1994; Hargreaves and Allen, 2003; Irmak et al., 2003; Itenfisu et al., 2003; Gavilán et al., 2006; Maeda et al., 2010). The use of these equations for the calibration of other empirical equations is justified by the lack of measured values of ET_0, as it is really difficult to find lysimeters at every weather station and, when available, low quality recorded data pose more errors than the introduction of low quality meteorological data in the equations (Irmak et al., 2003).

2.2.4 Spatial interpolation of daily ET estimates

The characterization of ET_0 at a regional scale is difficult due to the dependence of ET_0 with numerous factors. Two methodologies are commonly applied for the spatial interpolation of ET_0 in a regular matrix (McVicar et al., 2007): (1) interpolation of the input data, and, once all of them are available at cell scale, application of the model, also known as "interpolate and calculate" ($I - C$), or (2) calculation of ET_0 at each weather station from the input data, and then interpolation of the ET_0 results to the whole matrix, known as "calculate and interpolate" ($C - I$). When the Hargreaves equation is

used, the latter is immediate, but for the former, once the algorithms for the interpolation of meteorological variables have been developed, Hargreaves coefficient has to be spatially distributed. In order to do so, two methods were applied in this work: the interpolation of the adjusted coefficients at each station through inverse distance squared weighted, versus the allocation of the adjusted coefficients at each station to its region of influence.

The regionalization in the area in terms of evapotranspiration processes allows the definition of representative regions for hydrologically similar geographical units unlike other geometrical procedures (e.g. Thiessen polygons) that strictly depend on the spatial distribution of weather stations. In order to assess the region of local influence around each weather station, ET_0 values were estimated at each station by Hargreaves equation with the previously adjusted coefficients. Then, direct spatial interpolation through inverse distance squared weighted (IDW) was applied disregarding one station each time; the deviation of each calculation from the estimates considering all the stations was calculated in order to assess the relative influence of each station in the distributed computation. Thus, contours of influence were obtained by fixing a threshold of $\pm 0.1 \, \text{mm day}^{-1}$ for every station, differentiating between the wet and dry season. Finally, the adjusted coefficients at each station were assigned to its region of influence.

Finally, for the selection of the best interpolation method, three situations were compared in this study: the result of applying (1) IDW to ET_0 estimations at each station, a $C - I$ calculation type; and (2) distributed application of the Hargreaves equation with adjusted coefficients allocated by region, an $I - C$ calculation type; and (3) distributed application of Hargreaves equation with adjusted coefficients spatially interpolated by inverse distance squared weighted, an $I - C$ calculation type.

2.3 Validation

Finally, the validation of the adjusted coefficients was carried out with data recorded from 1 September 2008 to 2 July 2010. However, as for the surfaces of ET_0, the

uncertainty of the results depends on the goodness of fit of the results of the algorithms for the interpolation of each meteorological variable (Aguilar, 2008).

3   Results and discussion

3.1   Selection of the time step calculation in PM equation

Firstly, hourly estimates of ASCE-PM-ET_0 were calculated at the automatic weather stations with available hourly datasets (stations 601, 602, 603 and 802 in Fig. 1) and aggregated in order to obtain daily values. Fig. 2 represents aggregated hourly values (ET_0^{ASCE}) and daily ET_0 estimates calculated by ASCE-PM with the particularities of the daily time step (ET_0^{FAO}). A typical linear adjustment was applied where the zero intercept was forced through the origin and error estimates, the goodness of fit, R^2, as well as the slope of the adjustment, m, were computed with ET_0^{ASCE} estimates as the dependent variable (Table 2).

Although the results are quite conditioned by the quality of meteorological data, similar results to those obtained in previous studies were found: R between 0.95 and 0.97, which belongs to the range obtained by the ASCE (0.86 to 1.01 with 0.95 as annual mean (ASCE, 2000)) and similar to those obtained by Itenfisu et al. (2003) and Gavilán et al. (2008), the latter in a previous research at a regional scale in Andalusia. Similarly, the negative ME resulting values indicated a general underestimation at a daily scale, which appears to be due to the inability of the daily calculation to incorporate rough intra-daily variations in wind speed, vapour pressure or temperature, and also to the inclusion of the soil heat flux at an hourly scale in the energy balance, which is commonly estimated from other available variables, often from R_n (Allen et al., 1998; Gavilán et al., 2008).

In general, the close adjustment between both calculations at different time scales seems to be the result of different resistance coefficients between daytime and night time instead of the constant value suggested by Allen et al. (2006), and demonstrates

the internal consistency of the method at different time scales. However, the error obtained when using the daily-time step is not much higher than when using hourly-time steps (between −5 % at stations 603 and −3 % at station 802). Thus, as there are 3 more weather stations in the area that supply daily data and can help to define calibration criteria in the regional calibration of the Hargreaves coefficient (two in medium heights and one on the coast), these stations were included in the analysis, and so, the computation at a daily time scale was applied in the calibration process detailed hereafter.

3.2 Temporal variation of c_H

Daily ET_0 estimates obtained through the ASCE-PM method were compared to those calculated with the Hargreaves equation and different coefficients: the constant coefficient of 0.0023, and, then, adjusted c_H values at different temporal intervals.

Firstly, with the uncalibrated coefficient for the whole evaluation period, in general the Hargreaves equation overestimated the ET_0 values, as can be inferred from the mostly negative ME values found (Table 3), and only at stations 802 and 703 was a certain underestimation found. Thus, the resulting adjusted coefficients at every station except for 802 and 703 were lower than 0.0023 (c_H in Fig. 3). Therefore, the application of the adjusted coefficients at each station for the meteorological dataset of the evaluation period improved the ET_0^{HG} estimates in every case as the clouds of points were more aligned to the 1:1 slope line when compared to the daily estimates by ASCE-PM (Fig. 3), and, also, lower RMSE values than with the constant coefficient were always achieved as shown in Table 3.

Even though apparently there seems to be a relationship between the Hargreaves coefficient and elevation, according to Samani (2000) the value of c_H does not necessarily decrease with elevation, and he proposed a relationship between c_H and the daily temperature range. However, the different weather stations analysed showed a variable behaviour considering the adjusted coefficients and mean meteorological variables (Table 3), and so a distinction between coastal, inland and alpine stations was made.

At coastal stations, i.e. stations 603 and 701 in this study, the advective effects decrease the temperature range (ΔT in Table 3), which results in an enhanced underestimation of ET_0 by the Hargreaves equation, as the incoming solar radiation obtained through Eq. (3) is also underestimated (Samani, 2000; Gavilán et al., 2006; Jabloun and Sahli, 2008). This is the case of the above mentioned stations (603 and 701), where wind speed values are lower at the first station whereas the mean annual temperature range is slightly higher (Table 3), which means a lower coefficient at station 603 than at station 701 (c_H in Fig. 3).

At inland locations, i.e. stations 601, 602, 702 and 703 in this study, the temperature range is higher than at coastal stations, which would theoretically overestimate ET_0. However, the variability in the results found in previous studies suggests that many other factors affect the accuracy in ET_0 estimations such as cloudiness, humidity, topography, advection, proximity to water bodies, etc. In this way, similar trends to those obtained by Martínez-Cob and Tejero-Juste (2004) in a semiarid region were observed. Thus, from the results shown in Fig. 3, a constant value close to 0.0020 was obtained for semiarid inland locations with a monthly average wind speed lower than $2\,\mathrm{m\,s^{-1}}$, stations 702, 601 and 602 in the study area, whilst the original coefficient (0.0023) was at inland windy locations such as station 703.

The extreme case of the alpine station (802), with a considerable underestimation by the Hargreaves equation with the constant coefficient (ME $= 1.14\,\mathrm{mm\,day^{-1}}$ and RMSE $= 1.51\,\mathrm{mm\,day^{-1}}$), can be explained by the overestimation in the radiative over the aerodynamic term in the Penman equation (Eq. (1)). Previous studies carried out

at high altitudes found that this overestimation is not due to a lesser incoming net solar radiation in height, but to lesser storage of energy in the surface due to a thinner atmosphere than at sea level, that is, the lower Δ value due to the lower mean temperature (García et al., 2004), as shown in Table 3.

Following López-Urrea et al. (2006), who observed interannual patterns in ET_0 in semiarid areas due to the high temporal variability of meteorological conditions, the interannual calibration of c_H was applied at the seven stations in the study area. Results shown in Table 4 confirmed the global tendency already found: increasing values with the proximity to the sea at coastal stations, general underestimations at stations 703 and 802, and quite a constant value at inland locations of around 0.0020.

Intra-annual patterns were then evaluated in terms of the common wet (September–May) and dry (June–August) seasons in Mediterranean areas. When applying this differentiation between calibration periods, higher coefficients were obtained in the dry season at every station, with a greater oscillation at coastal stations where values up to 0.0015 for the wet and 0.0025 for the dry season were found, respectively (Table 5). The classical seasonal distinction was also performed (Aguilar, 2008), and similar trends were obtained as the highest values were reached at every station every year in spring and summer, especially in spring, close to 0.0023, whilst in winter and autumn, the coefficient values dropped by up to 0.0012 (in autumn). Again, the wider range of variability was found at coastal stations, and at station 802 the highest values were found in winter and autumn.

In order to select a set of adjusted coefficients, a simple comparison of error values among the different possibilities was applied. From the contrast of results, error values improved in every station when the computation was applied with variable coefficients between the wet and the dry season (Table 5) over those obtained with adjusted coefficients for the whole evaluation period, or with the uncalibrated coefficient (Table 3), especially at stations 603 and 802, where RMSE values with the 0.0023 constant coefficient were higher than $1\,\mathrm{mm\,day^{-1}}$.

Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

◄◄ | ►►

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

From these results, a variable coefficient between the wet and the dry season was proposed for the subsequent spatial interpolation.

The calibrated coefficients were validated with available datasets registered from 1 September 2008 to 2 July 2010. Again, ME and RMSE values were used to quantify the differences between the ET_0 estimated using the reference method (ET_0^{FAO}) and the estimates obtained using the Hargreaves equation with the 0.0023 constant coefficient and the adjusted coefficients for wet and dry seasons (ET_0^{HG}). The application of seasonally adjusted coefficients improved the estimates with the Hargreaves equation as shown in Fig. 4, where the cloud of points tends to approximate more than Hargreaves estimates with the constant coefficient to the 1:1 slope line at every weather station. The error values obtained in the validation process are summarized in Table 6. It can be seen that the order of magnitude of the error estimates remains similar to the ones obtained in the evaluation process (Tables 3 and 5). Furthermore, the lower error values obtained with the calibrated coefficient (Table 6) validate their use in the study area instead of the 0.0023 constant value.

3.3 Spatial interpolation of c_H

Once the threshold values of influence were fixed as previously exposed, differentiating between the wet and dry seasons, the different regions of influence of each station can be seen in Fig. 5, together with the distributed ET_0^{FAO} estimates accumulated for the wet and the dry seasons per hydrological year throughout the evaluation period (2004–2008). The Southern hillside of Sierra Nevada, and, in general, alpine regions, are included in the region of station 802, while along the seaside both coastal stations, station 701 and 603, share their influence.

The locally adjusted seasonal coefficients (wet and dry season) as justified in the previous section (Table 5) were assigned to the region of influence of each station. In order to evaluate the efficiency of each alternative, the deviation to ASCE-PM ET_0 estimates at cell scale were computed, and simple statistics of the deviation in terms of

the cell by cell difference between the reference and the empirical methods were calculated for each alternative. Figure 6 shows the difference between ET_0 estimations by ASCE-PM equation (Fig. 5) and ET_0 estimations by Hargreaves equation with the adjusted coefficients interpolated by IDW. Here, the main under- and overestimations take place mainly in the dry season in the alpine areas as well as throughout the main river valleys. However, when doing the same computation with the adjusted coefficients allocated to regions of influence, extreme differences were observed such as underestimations in the alpine regions that are captured by stations 702 and 703's area of influence, and overestimations in deep valleys that are captured by station 802's region. Thus, a modification to the region assigned to station 802 was proposed. For this, the boundary of the area of influence of station 802 was replaced by the curve of 2000 m altitude as this appears to be the limit of the linear trend of rainfall with height in the area (Moñino et al., 2011), keeping the rest of the regions as originally drawn (Fig. 7).

Table 7 shows the statistics of the difference between ASCE-PM ET_0 estimates at cell scale and the different alternatives. In general, IDW interpolation of the Hargreaves equation at each station with adjusted c_H (a in Table 7) always gave the highest mean values of the deviation. Thus, the average of mean deviations was $-0.06\,\mathrm{mm\,day}^{-1}$ and $-0.33\,\mathrm{mm\,day}^{-1}$ for the wet and dry seasons respectively, which accounts for mean differences of $-20.75\,\mathrm{mm\,(wet\ season}^{-1})$ and $-30\,\mathrm{mm\,(dry\ season}^{-1})$. In this way, $50.75\,\mathrm{mm\,year}^{-1}$ in an area with a mean annual rainfall of $500\,\mathrm{mm\,year}^{-1}$ can constitute a considerable source of error in all the hydrological processes where evapotranspiration processes are involved. Therefore, the suitability of applying $I-C$ methods rather than $C-I$ methods in complex terrain watersheds is demonstrated. In the three $I-C$ methods applied in this study, the lowest mean values of the deviation were obtained with the distributed computation of the Hargreaves equation with the adjusted c_H distributed according to the regionalization shown in Fig. 7 (c in Table 7) (average of the mean deviations of $0.02\,\mathrm{mm\,day}^{-1}$ and $0.15\,\mathrm{mm\,day}^{-1}$ for the wet and dry seasons, respectively, which accounts for mean differences of $3.38\,\mathrm{mm\,(wet\ season}^{-1})$

and 13.43 mm (dry season^{-1})). Nevertheless, the order of magnitude of the statistics for the three $I-C$ methods applied in this study (b, c and d in Table 7) suggests that further analyses must be made in order to obtain smoothed (more realistic) gradients along the contour of the 802's area of influence as well as throughout the valleys in-
5 cluded in the region of station 702's area of influence, whose current values produce the higher standard deviation values associated with the use of regions of influence in the interpolation procedure (b and c in Table 7).

4 Conclusions

The interplay of different factors at different temporal scales became evident through
10 the results shown. Thus, the calculation of ET$_0$ must be made dependent on the temporal scale at which the water and energy balance is applied in a model. At an hourly scale, ASCE-PM's equation is undoubtedly the most suitable method provided that all the necessary input data are available at this temporal scale. However, even though the same equation can be applied at a daily scale, the satisfactory results obtained by
15 the Hargreaves equation at this temporal scale, once the coefficient has been locally calibrated, situate this method as an alternative to ASCE-PM's method for situations where not all the input data are available, or where distributed calculations must be performed over long periods/large areas.

The application of PM's equation at different time scales proved the method's inter-
20 nal consistency at the four stations considered, which included a coastal, two inland and one alpine region, and confirms its use as reference value. Here, the linear adjustment between daily estimates and aggregated hourly values, with the zero intercept forced through the origin, resulted in R^2 values higher than 0.99 at the four stations considered.

25 The Hargreaves equation was locally calibrated with satisfactory results in the study site. With the original parameter of 0.0023, the Hargreaves equation overestimated ET$_0$ at every station except at station 703 and at the alpine station 802, with RMSE

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

values with respect to ASCE-PM's daily estimates close to $1\,\text{mm}\,\text{day}^{-1}$ and up to $1.51\,\text{mm}\,\text{day}^{-1}$ at the alpine station. At coastal sites, underestimation increased as the temperature range decreased. The local adjustment for the whole period showed different tendencies depending on the proximity to the seaside. At coastal stations,
5 a higher coefficient as the temperature decreases is proposed. For inland stations, a constant value of around 0.0020 is proposed for non-windy locations, whilst the original coefficient (0.0023) is maintained for windy locations. Finally, at the alpine station, the considerable underestimation must be compensated by increasing the coefficient up to 0.0038. The same trends were obtained when the calibration was carried out on a hy-
10 drological year temporal basis. However, when differentiating between the wet season and the dry season in the calibration process, higher coefficients were obtained at every station in the dry season with a greater variation at coastal (from 0.0015 to 0.0025) and non windy inland locations (from 0.0017 to 0.0021), which improved in every case the error values found, especially at stations 603 and 802, where RMSE values with
15 the 0.0023 constant coefficient were higher than $1\,\text{mm}\,\text{day}^{-1}$. These higher deviations found when a constant coefficient is used can imply considerable errors that further affect the calculation of the soil drying and snow dynamics, among others.

Finally, the four spatial interpolation methods applied allowed the application of the Hargreaves equation at a daily scale with few deviations from the calculation through
20 the equation of the ASCE-PM, lower than $\pm 1\,\text{mm}\,\text{day}^{-1}$ in the extreme values of all the examples analysed. However, the highest deviations obtained when applying the $C-I$ method (close to $50.75\,\text{mm}\,\text{year}^{-1}$) demonstrated the suitability of applying $I-C$ methods rather than $C-I$ methods in these complex terrain watersheds. The lowest deviations (close to $17\,\text{mm}\,\text{year}^{-1}$) were obtained when distributing Hargreaves equation
25 with locally adjusted Hargreaves coefficients for each weather station and its associated region differentiating between the wet and the dry season. A further analysis is proposed in order to reduce the resulting gradients in the low height valleys along the river as well as in the contour of the alpine station's region of influence.

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Title Page

Abstract | Introduction

Conclusions | References

Tables | Figures

◄ | ►

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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Calculation of reference evapotranspiration surfaces

C. Aguilar and M. J. Polo

Title Page

Abstract	Introduction
Conclusions	References
Tables	Figures

|◄ | ►|

◄ | ►

Back | Close

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

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Table 1. Resistance coefficient values in ASCE-PM equation (Gavilán et al., 2007).

	Daily	Hourly
C_n	900	37
C_d	0.34	0.24 (daytime)/0.96 (night time)

Table 2. Mean error (ME ($\mathrm{mm\,day^{-1}}$)), mean absolute error (MAE ($\mathrm{mm\,day^{-1}}$)) and root mean squared error (RMSE ($\mathrm{mm\,day^{-1}}$)), ratio between mean estimations (R), slope (m) and goodness of the linear fit (R^2) between $\mathrm{ET_0^{FAO}}$ (independent variable) and $\mathrm{ET_0^{ASCE}}$ (dependent variable).

Station	m	R^2	ME	MAE	RMSE	R
601	0.982	0.992	−0.122	0.161	0.196	0.96
602	0.981	0.991	−0.135	0.174	0.213	0.96
603	0.977	0.991	−0.122	0.163	0.197	0.95
802	0.959	0.990	−0.131	0.180	0.223	0.97

Table 3. Mean error (ME (mm day^{-1})) and root mean squared error (RMSE (mm day^{-1})) with both the constant 0.0023 coefficient and the global adjusted c_H, and mean values for the evaluation period (20/11/2004–31/08/2008) of temperature (T), wind speed (v) and relative humidity (HR).

Station	z(m)	$c_H = 0.0023$		Global adjusted c_H		$\bar{T}(°C)$	$\Delta T(°C)$	$\bar{v}(\text{m s}^{-1})$	HR(%)
		ME	RMSE	ME	RMSE				
701	35	−0.39	0.91	−0.20	0.89	18.85	7.72	0.96	62
603	49	−0.87	1.15	−0.24	0.95	19.91	8.57	0.80	65
703	400	0.05	0.88	−0.05	0.87	17.60	12.08	2.11	60
702	700	−0.71	0.99	−0.24	0.83	14.40	10.81	1.38	59
602	781	−0.63	0.93	−0.14	0.75	15.36	12.17	1.15	55
601	950	−0.65	0.94	−0.20	0.80	15.18	9.68	1.5	54
802	2510	1.14	1.51	−0.06	0.74	6.26	5.89	3.05	24

Table 4. Adjusted c_H values for each hydrological year (1 September to 31 August).

Station	z(m)	Annual c_H				
		2004–2005	2005–2006	2006–2007	2007–2008	Average
701	35	0.002169	0.002032	0.002130	0.002318	0.002162
603	49	0.001846	0.001757	0.001848	0.001974	0.001856
703	400	0.002447	0.002290	0.002302	0.002400	0.002360
702	700	0.002172	0.001937	0.001894	0.001919	0.001981
602	781	0.002162	0.002008	0.001905	0.001919	0.002001
601	950	0.001936	0.001959	0.002037	0.002033	0.001991
802	2510	0.004101	0.003741	0.003648	0.003690	0.003795

Table 5. Mean error (ME (mm day^{-1})) and root mean squared error (RMSE (mm day^{-1})) with the c_H adjusted by wet/dry season for the evaluation period (20/11/2004–31/08/2008).

Station	z(m)	c_H Dry season	Wet season	ME	RMSE
701	35	0.002520	0.001842	−0.10	0.74
603	49	0.002220	0.001503	−0.12	0.77
703	400	0.002415	0.002298	−0.02	0.86
702	700	0.002192	0.001721	−0.14	0.73
602	781	0.002102	0.001874	−0.09	0.72
601	950	0.002166	0.001780	−0.13	0.74
802	2510	0.003960	0.003591	−0.03	0.72

Table 6. Mean error (ME (mm day^{-1})) and root mean squared error (RMSE (mm day^{-1})) for the validation period (01/09/2008–02/07/2010) with the constant 0.0023 coefficient and the c_H adjusted by wet/dry season.

Station	z(m)	$c_H = 0.0023$		Wet/dry season adjusted c_H	
		ME	RMSE	ME	RMSE
701	35	−0.26	0.74	0.10	0.72
603	49	−0.77	1.05	0.03	0.78
703	400	−0.37	0.76	−0.42	0.76
702	700	−0.65	0.89	−0.06	0.67
602	781	−0.50	0.84	0.01	0.71
601	950	−0.41	0.79	0.10	0.75
802	2510	1.26	1.53	0.25	0.57

Full Screen / Esc

Table 7. Mean (μ (mm day^{-1})) and standard deviation (σ (mm day^{-1})) of the difference between ASCE-PM ET$_0$ estimates and different alternatives per hydrological: **(a)** IDW to Hargreaves estimates at each station with adjusted c_H, ($C - I$ method type) **(b)** distributed computation of Hargreaves equation with adjusted c_H allocated to regions of influence **(c)** distributed computation of Hargreaves equation with adjusted c_H allocated to modified regions of influence **(d)** distributed computation of Hargreaves equation with IDW to adjusted c_H.

Wet Season	04/05		05/06		06/07		07/08		08/09		09/10	
	μ	σ	μ	σ	μ	σ	μ	σ	μ	σ	μ	σ
a	0.02	0.35	−0.05	0.34	−0.09	0.34	−0.06	0.36	−0.09	0.32	−0.09	0.31
b	0.01	0.44	0.05	0.42	0.02	0.41	0.03	0.44	0.05	0.40	0.06	0.39
c	0.06	0.42	0.02	0.39	−0.02	0.39	0.002	0.41	0.03	0.38	0.03	0.36
d	0.07	0.38	0.03	0.35	−0.02	0.34	0.006	0.37	0.033	0.32	0.04	0.32

Dry Season	04/05		05/06		06/07		07/08		08/09			
	μ	σ	μ	σ	μ	σ	μ	σ	μ	σ		
a	−0.21	0.44	−0.49	0.57	−0.41	0.50	−0.28	0.41	−0.24	0.40		
b	−0.10	0.64	−0.29	0.68	−0.30	0.69	−0.18	0.64	−0.16	0.65		
c	−0.11	0.59	−0.30	0.64	−0.30	0.65	−0.19	0.61	0.17	0.61		
d	−0.15	0.53	−0.34	0.55	−0.34	0.53	−0.23	0.49	−0.21	0.51		

C. Aguilar and M. J. Polo
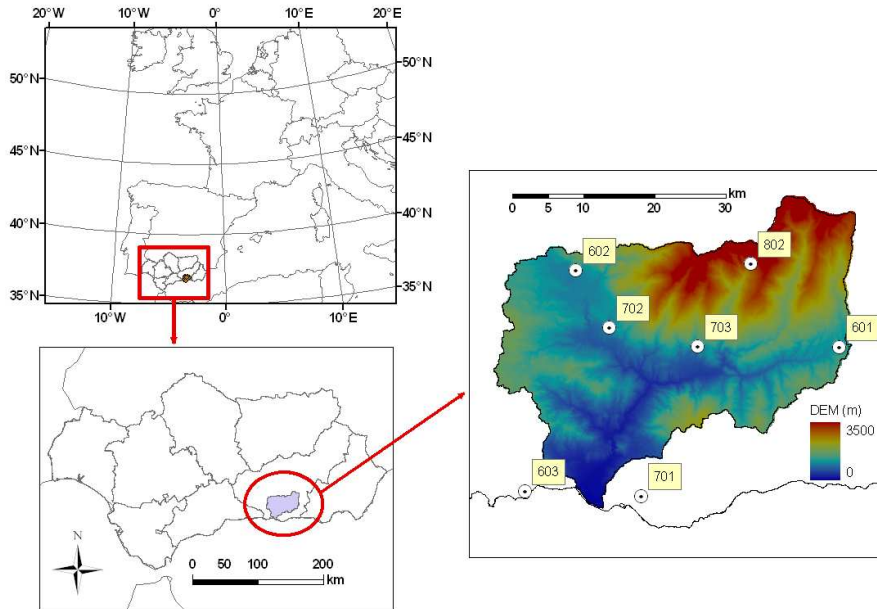
Fig. 1. Study site location, and weather stations and DEM.

Fig. 2. ET_0^{ASCE} vs. ET_0^{FAO} at the four selected weather stations for the evaluation period (20/11/2004–31/08/2008).

Full Screen / Esc

Printer-friendly Version

Interactive Discussion

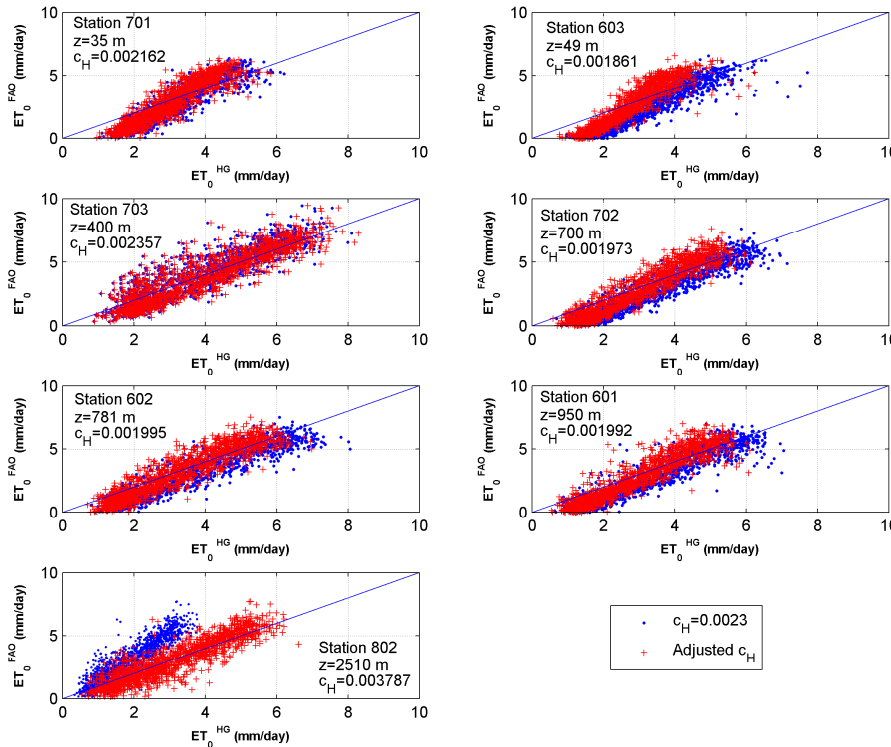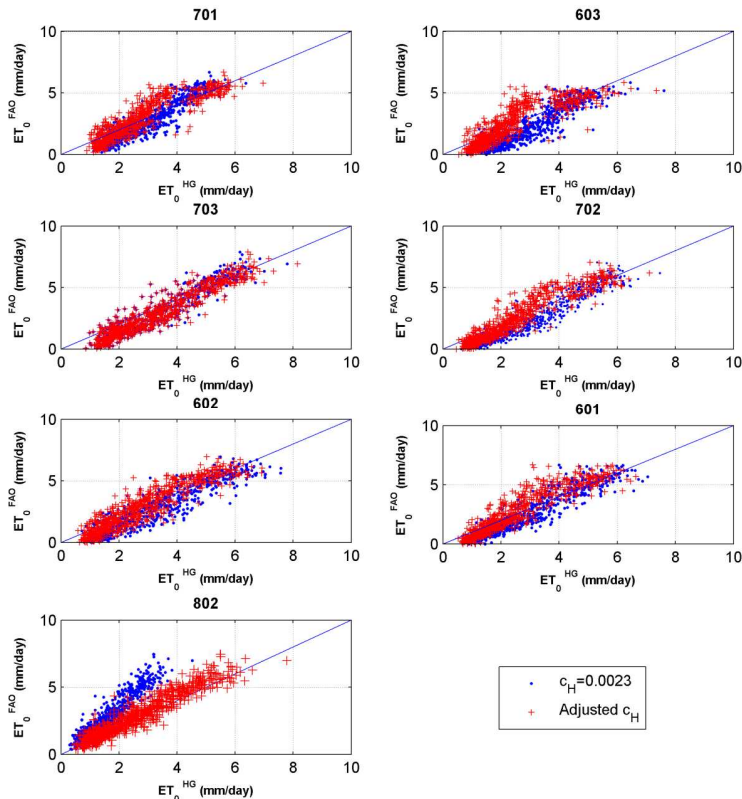**Calculation of reference evapotranspiration surfaces**

C. Aguilar and M. J. Polo

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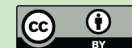
Fig. 3. Daily ET_0(ET_0^{FAO}) vs. Daily ET_0 by Hargreaves (ET_0^{HG}) with a constant coefficient (0.0023) and adjusted c_H at each station for the evaluation period (20/11/2004–31/08/2008).

Fig. 4. Daily ET_0 (ET_0^{FAO}) vs. Daily ET_0 by Hargreaves (ET_0^{HG}) with a constant coefficient (0.0023) and adjusted wet/dry season c_H at each station for the validation period (01/09/2008–02/07/2010).
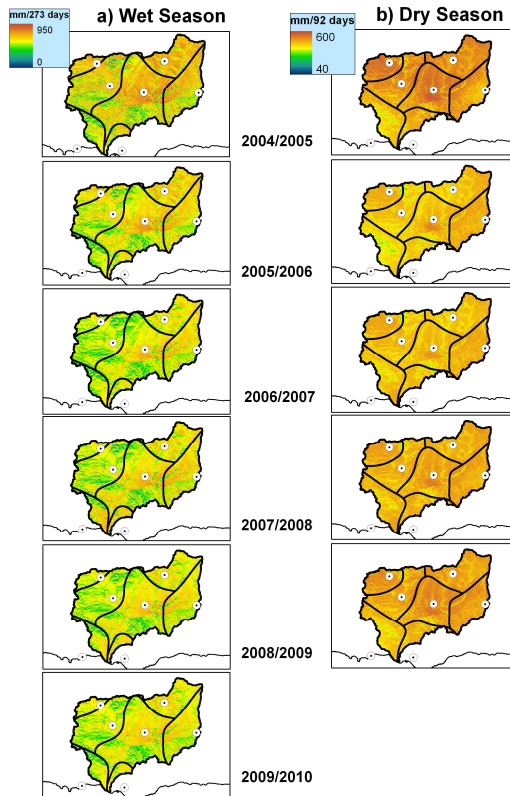
Fig. 5. Regions of influence and ASCE-PM ET$_o$ estimates for the **(a)** wet season (01/09–31/05) and **(b)** the dry season (01/06–31/08).
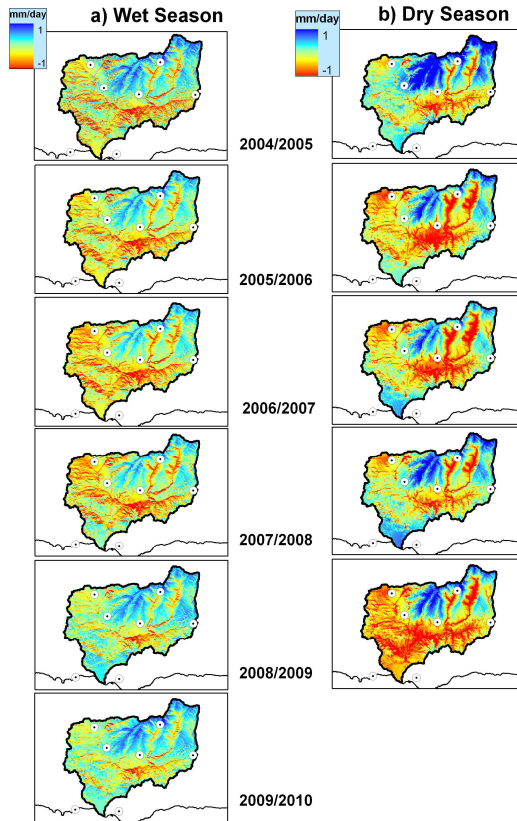
Fig. 6. Differences between ET$_o$ obtained by ASCE-PM and Hargreaves with adjusted coefficient interpolated by IDW.
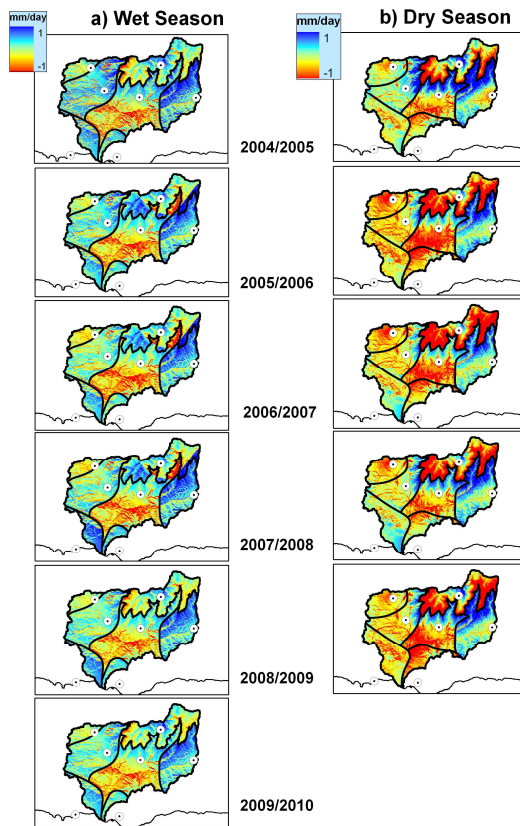
Fig. 7. Modified regions of influence and Hargreaves and differences between ET_o obtained by ASCE-PM and Hargreaves with adjusted coefficient allocated to modified regions of influence.