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ORIGINAL PAPER

Daily reference crop evapotranspiration with reduced data sets in the humid environments of Azores islands using estimates of actual vapor pressure, solar radiation, and wind speed

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Abstract

Reference crop evapotranspiration (ET_o) estimations using the FAO Penman-Monteith equation (PM-ET_o) require a set of weather data including maximum and minimum air temperatures (T_{max}, T_{min}), actual vapor pressure (e_a), solar radiation (R_s), and wind speed (u_2). However, those data are often not available, or data sets are incomplete due to missing values. A set of procedures were proposed in FAO56 (Allen et al. 1998) to overcome these limitations, and which accuracy for estimating daily ET_o in the humid climate of Azores islands is assessed in this study. Results show that after locally and seasonally calibrating the temperature adjustment factor a_d used for dew point temperature (T_{dew}) computation from mean temperature, ET_o estimations shown small bias and small RMSE ranging from 0.15 to 0.53 mm day^{-1}. When R_s data are missing, their estimation from the temperature difference (T_{max}–T_{min}), using a locally and seasonal calibrated radiation adjustment coefficient (k_{Rs}), yielded highly accurate ET_o estimates, with RMSE averaging 0.41 mm day^{-1} and ranging from 0.33 to 0.58 mm day^{-1}. If wind speed observations are missing, the use of the default $u_2 = 2$ m s^{-1}, or 3 m s^{-1} in case of weather measurements over clipped grass in airports, revealed appropriated even for the windy locations ($u_2 > 4$ m s^{-1}), with RMSE < 0.36 mm day^{-1}. The appropriateness of procedure to estimating the missing values of e_a, R_s, and u_2 was confirmed.

Keywords PM-ET_o equation · Actual vapor pressure from temperature · Solar radiation from temperature · Default wind speed

1 Introduction

The reference crop evapotranspiration (ET_o), often named potential ET in hydrologic and climatologic studies, is a main factor for characterizing the local climate and for computing crop and vegetation evapotranspiration. Its accurate estimation is relevant in several domains such as crop water management, irrigation planning and management, hydrologic and water balance studies, climate characterization, and climate change analysis as revised by Pereira et al. (2015) and Pereira (2017). This is the case of the Azores islands, where climate studies with model CIELO (Azevedo et al. 1999) would benefit if complemented with ET_o information when only reduced data sets are available as discussed below in Section 2.1.

Grass reference ET_o was defined by FAO after parameterizing the Penman-Monteith equation (Monteith 1965) for a cool season grass (Smith et al. 1991). ET_o was then defined as the rate of evapotranspiration from a hypothetical reference crop with an assumed crop height $h = 0.12$ m, a fixed daily canopy resistance $r_s = 70$ s m^{-1}, and an albedo of 0.23, closely

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resembling the evapotranspiration from an extensive surface of green grass of uniform height, actively growing, completely shading the ground, and not short of water (Allen et al. 1998). Later studies confirmed the appropriateness of using a daily fixed surface resistance of 70 s m^{-1} (Lecina et al. 2003; Steduto et al. 2003). This definition allowed a clear distinction relative to the Penman's potential ET (Penman 1948) and the equilibrium evaporation (Slatyer and McIlroy 1961), thus assuming a well-defined relationship between ET_o and actual ET of a vegetation surface (Pereira et al. 1999).This relationship bases upon the aerodynamic and surface resistances of the reference grass crop and of the considered vegetation, thus theoretically supporting the concept and use of crop coefficients (Pereira et al. 1999). The Penman-Monteith ET_o (PM-ET_o) is therefore defined by an equation derived from the Penman-Monteith combination equation when parameterized as referred above (Allen et al. 1998). Daily ET_o (mm day^{-1}) is therefore given as

$$\mathrm{ET_o} = \frac{0.408\Delta(R_n - G) + \gamma\dfrac{900}{T_{mean}+273}u_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)} \quad (1)$$

where R_n is the net radiation at the crop surface (MJ m^{-2} day^{-1}), G is the soil heat flux density (MJ m^{-2} day^{-1}), T_{mean} is the mean daily air temperature at 2 m height (°C), u_2 is the wind speed at 2 m height (m s^{-1}), e_s is the saturation vapor pressure at 2 m height (kPa), e_a is the actual vapor pressure at 2 m height (kPa), VPD or e_s–e_a is the saturation vapor pressure deficit (kPa), Δ is the slope vapor pressure curve (kPa °C^{-1}), and γ is the psychrometric constant (kPa °C^{-1}). For daily computations, G equals zero as the magnitude of daily soil heat flux beneath the grass reference surface is very small (Allen et al. 1998). Computations therefore require observed data on maximum and minimum temperature (T_{max} and T_{min}), solar radiation (R_s) or sunshine duration (n), air relative humidity (RH) or psychrometric data, and wind speed (u_2). While T_{max} and T_{min} records are commonly well observed at weather station networks in most countries, the other variables are often not easily available with good quality, or are available for only short periods of time, and/or data sets have frequent gaps; in addition, their acquisition may be very expensive.

To face conditions of incomplete/reduced data sets, methods were proposed by Allen et al. (1998) for estimation of the missing variables R_s, e_a, and/or u_2. When relative humidity data or psychrometric observations are missing, Allen et al. (1998), following a previous assessment by Jensen et al. (1997), recommended to estimate actual vapor pressure (e_a) assuming that T_{min} could be an acceptable estimate of dew point temperature (T_{dew}):

$$e_a = e^o(T_{dew}) = 0.611 \exp\left[\frac{17.27\,T_{min}}{T_{min} + 237.3}\right] \quad (2)$$

Nevertheless, Allen et al. (1998) referred the need to correct weather data measured in weather stations not respecting the assumption that collected data refer to vertical fluxes of heat and water vapor originated by an extended green grass vegetated surface (Allen 1996). In addition, Kimball et al. (1997) observed that (a) T_{min} exceeds T_{dew} in arid sites leading to average daily differences of 0.8 to 1.2 kPa between e_a computed from T_{min} and T_{dew}; (b) in sites with semiarid climate there, is seasonality in the differences between T_{min} and T_{dew} resulting seasonal differences between respective e_a values, varying from 0.1 to 0.6 kPa for winter and summer months, respectively; (c) smaller differences between T_{min} and T_{dew} occurred in other less arid climates, e.g., coastal Mediterranean, humid continental, and humid subtropical conditions, with average daily differences of less than 0.3 kPa between e_a computed from T_{min} and T_{dew}. The need to correct T_{min} to estimate T_{dew} in arid areas was well identified (Temesgen et al. 1999) but, for a decade, literature has been reporting that users were assuming $T_{dew} = T_{min}$. This may be due to the fact that, with few exceptions, researchers were focusing on alternative ET equations or numerical methods that do not require computation of e_a, or because Allen et al. (1998) did not propose a well-defined procedure but to perform corrections to T_{min} after results of data analysis. Later, studies by Todorovic et al. (2013), Raziei and Pereira (2013) and Ren et al. (2016) developed a clear set of corrections for T_{min} relative to various arid climatic conditions and corrections for the mean daily temperature for estimating T_{dew} in sub-humid and humid climates where T_{dew} is likely above T_{min}. These corrections were set according to the UNEP aridity index (AI; UNEP 1997), which is defined by the ratio between the mean annual precipitation (P) and the mean annual potential climatic evapotranspiration (PET_{TH}, Thornthwaite 1948).

Corrections for aridity consist of

$$T_{dew} = T_{min} - a_T \quad (3)$$

where the correction factor a_T varies with the climate aridity of the station:

a) Hyper-arid locations, AI < 0.05, $a_T = 4$ °C,
b) Arid locations, AI from 0.05 to 0.20, $a_T = 2$ °C,
c) Semiarid locations, AI from 0.20 to 0.50, $a_T = 1$ °C,
d) Dry sub-humid locations, AI from 0.50 to 0.65, $a_T = 1$ °C.

Considering the relations for T_{dew} in moist air proposed by Lawrence (2005), for moist sub-humid and humid climates and when the mean temperature is very low, T_{dew} was empirically approximated (Todorovic et al. 2013) by

$$T_{dew} = T_{mean} - a_d \quad (4)$$

where $a_d = 2$ °C when $0.8 < P/PET_{TH} < 1.0$ and $a_d = 1$ °C if $P/PET_{TH} > 1.0$. However, later studies (Ren et al. 2016) have shown that a_d varies with factors other than aridity and, depending upon the accuracy requirements, its calibration is needed. It results the equation

$$e_a = e^o(T_{dew}) = 0.611 \exp\left[\frac{17.27\,(T_{mean} - a_d)}{(T_{mean} - a_d) + 237.3}\right] \quad (5)$$

where T_{dew} is replaced by its value in Eq. 4.

Allen (1997) and Allen et al. (1998) proposed to estimate R_s for using with the PM-ET$_o$ equation (Eq. 1) adopting the R_s predictive equation by Hargreaves and Samani (1982) that expresses R_s as a linear function of the square root of the temperature difference $T_{max} - T_{min}$:

$$R_s = k_{R_s}(T_{max} - T_{min})^{0.5} R_a \quad (6)$$

where k_{Rs} is an empirical radiation adjustment coefficient (°C$^{-0.5}$) and R_a is the extraterrestrial radiation (MJ m^2 day^{-1}). Comparative studies by Abraha and Savage (2008), Bandyopadhyay et al. (2008) and Aladenola and Madramootoo (2014) reported good performance of the temperature difference Eq. 6 relative to other models.

The empirical coefficient k_{Rs} was originally considered to range 0.16 to 0.19 °C$^{-0.5}$, respectively for "interior" or "coastal" regions (Allen 1997; Allen et al. 1998).However, k_{Rs} is supposed to vary with altitude, thus reflecting the air pressure changes as for the volumetric heat capacity of the atmosphere (Allen 1997), which increases linearly with the volumetric mass density and thus the atmospheric pressure (Turbet et al. 2017). Thus, k_{Rs} would vary spatially not only with site elevation but also with the distance to sea as earlier discussed by Allen et al. (1998) and is compatible with influences of air moisture on the volumetric heat capacity of the atmosphere. In addition, k_{Rs} was observed to increase with the aridity of the site and with wind speed (Raziei and Pereira 2013; Ren et al. 2016). Other factors may also influence k_{Rs}. Samani (2000) developed a quadratic relationship between k_{Rs} and the temperature difference $T_{max} - T_{min}$, but this approach does not contribute to explain the spatial variability of k_{Rs}. The effect of altitude was considered in the relationship between k_{Rs} and the square root of the ratio between the atmospheric pressure at the location and at sea level (Allen 1997). Annandale et al. (2002) also developed a decreasing relationship between k_{Rs} and elevation to account for the effects of reduced atmospheric thickness on R_s. Therefore, for island locations, where climate is always influenced by the sea proximity, it is required to calibrate k_{Rs} and not assume the above referred range of its values, respectively 0.16 to 0.19 °C$^{-0.5}$ for "interior" or "coastal" regions.

In the original version of the Hargreaves-Samani (HS) equation (Hargreaves and Samani, 1982), a bulk constant term of 0.023 is used, which is known as the Hargreaves coefficient and corresponds to the product 0.0135 k_{Rs}, with k_{Rs} = 0.17 °C$^{-0.5}$, and 0.0135 representing a units' conversion constant. Various researchers calibrated that coefficient, i.e., inexplicitly calibrated k_{Rs}, using a two-parameter temperature function (Vanderlinden et al. 2004; Mendicino and Senatore 2013; Martí et al. 2015). However, results by these authors show that using two-parameter temperature function to calibrate the Hargreaves coefficient produced quite different parameters for various locations having similar climate in the Mediterranean area, which makes it difficult to interpret results and does not allow to identify trends for these parameters. A function with a single parameter focusing on the Hargreaves coefficient have shown that given the variety of factors influencing R_s and $T_{max} - T_{min}$, there is the need for searching the best value for k_{Rs} at each location.

Allen et al. (1998) proposed the use of the world average wind speed value u_2 = 2 m s^{-1} as the default estimator of wind speed when related data are missing. When average local wind speed data are available, they were used alternatively following the assessments by Popova et al. (2006) and Jabloun and Sahli (2008).

The use of the referred approaches for estimating the parameters of the PM-ET$_o$ equation allows computing ET$_o$ just replacing R_s, e_a, or u_2 by their respective estimators when related weather variables are missing, or using temperature data only when more variables are missing. Considering the discussion above on procedures to compute ET$_o$ with reduced data sets when estimating just the missing variables, as well as the lack of appropriate information on estimating ET$_o$ and related missing variables in humid island environments, the objective of this study consists of assessing the performance of temperature-based procedures to estimate actual vapor pressure (Eqs. 2 through 5), solar radiation (Eq. 6), and wind speed when used in PM-ET$_o$ computations in humid island environments. The assessment of ET$_o$ computed with only temperature data is the object of a companion paper (Paredes et al. 2017).

2 Materials and methods

2.1 Study area and data

The archipelago of Azores, of volcanic origin, comprises nine islands and is located in the North Atlantic at latitudes 36° 45′ N to 39° 43′ N and longitudes 24° 45′ W to 31° 17′ W (Fig. 1). The east-most island, Santa Maria, is located approximately 1400 km from mainland Europe, and the west-most island, Flores, lies 1900 km from the North American continent (Fig. 1). Biogeographically, the archipelago of Azores belongs to Macaronesia, which also comprises the archipelagos of Madeira, Canaries, and Cape Verde (Cropper and Hanna 2014), which are drier or much drier than Azores.

Daily data used in the present study was collected in various weather stations in eight islands whose locations, basic

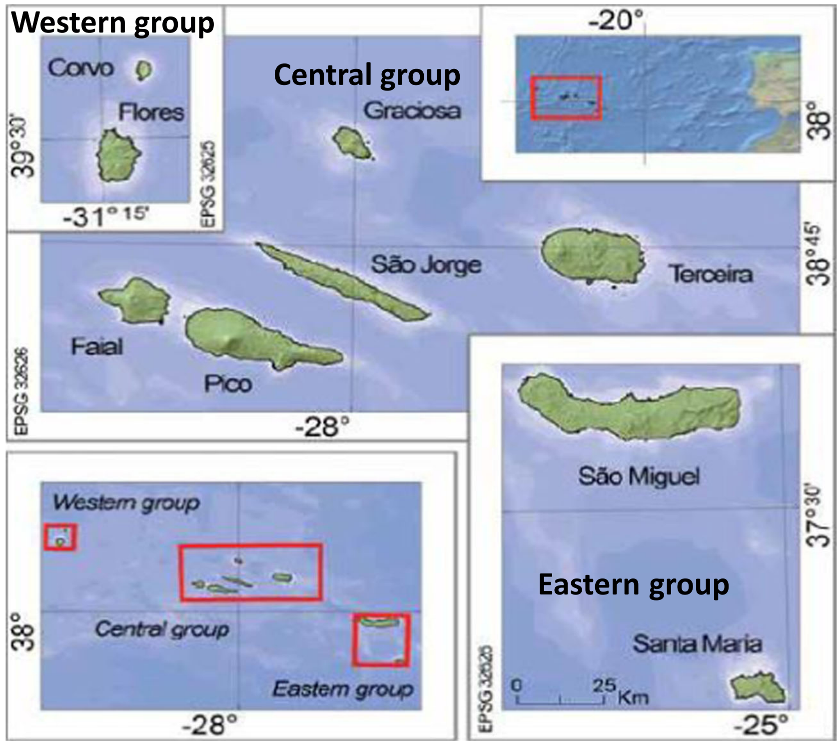

Fig. 1 The Azores archipelago and location of islands under study (adapted from Elias et al. 2016)

climate characteristics, and size of data sets are described in Tables 1 and 2. Data included precipitation (P), maximum and minimum air temperatures (T_{max} and T_{min}, °C), relative humidity (RH, %), wind speed (u_2, m s^{-1}), and solar radiation (R_s, W m^{-2}) or sunshine duration (n, h). All data were collected above green grass, including those collected in airports. The basic information about variability of daily collected data by month is given in Fig. 2 for eight selected stations, which refer to seven out of the nine islands and two contrasting environments in Terceira island.

The Azorean climate is strongly influenced by its location in the middle of the North Atlantic Ocean. During most of the year (September to March), the Azores region is frequently crossed by the North Atlantic storm-track, the main path of rain-producing weather systems, while during late Spring and Summer, the Azores climate is influenced by the Azores anticyclone (Santos et al. 2004). Therefore, the yearly average precipitation varies with longitude, from 730 mm year^{-1} in Santa Maria, the east-most island, to 1666 mm year^{-1} in Flores, the west-most island (Barceló and Nunes 2012).

As depicted in Fig. 2, the weather variables vary much, not only within the year but also within the months, which is a characteristic of the islands' climate environment as referred above. Precipitation is more abundant during November to January and varies much within each month, with elevation and longitude. Air humidity is generally high, with median RH values close to 80% and RH_{min} near 60%. The monthly average temperature varies little along the year, varies much within a month, and the cooler months are generally January and February and the hottest one is August. Frosts are rare below 600 m of altitude. The dominant winds are from SW, with high moisture; winds are quite strong and show a great variability within months mainly in winter. Solar radiation is highest generally by July and smaller by December; its variability is very high mainly in summer months. Therefore, the reference evapotranspiration ET_o is greater in summer months when the variability within months is quite high. According to the Köppen-Geiger classification (Kottek et al. 2006), the climate in Faial and Flores is Cfb, fully humid with warm summer (Table 1); in Pico, Terceira, and the western part of S. Miguel (Lagoa do Fogo, Furnas, and Tronqueira) is Csb, humid with dry and warm summer, and in Graciosa, S. Jorge, Santa Maria, and the eastern part of S. Miguel (Chã de Macela, Ponta Delgada, Santana, and Sete Cidades) is Csa, humid with a dry and hot summer. In all locations, the aridity ratio P/PET_{TH} is much higher than 1.0, from 2.0 at Santa Cruz da Graciosa to 8.6 at S. Caetano (Table 1), indicating a very humid climate (UNEP, 1997).

Several studies have been performed relative to the climate of the Azores islands using the CIELO model (Azevedo et al. 1999), which is a physically based model that simulates the transformations experienced by an humid air mass ascending from the windward side, starting from the sea level up to the central mountains of the islands, and then descending from the opposite side. During the ascending path, the air column's temperature decreases and the air saturates, with the consequent condensation of water vapor that precipitates in favorable physical conditions; on the

Table 1 Weather stations characterization and data range of the weather data series

Weather stations	Island	Latitude (N)	Longitude (W)	Elevation (m)	P/PET*	Köppen classe	Observations	
							Dates	Number of days
Santa Cruz das Flores**	Flores	39° 27′ 31.11″	31° 07′ 49.65″	28	4.9	Cfb	2003–13	3539
Horta**	Faial	38° 31′ 45.31″	28° 37′ 43.20″	40	3.4	Cfb	2006–10	3349
São Caetano	Pico	38° 27′ 0.3″	28° 26′ 2.42″	770	8.6	Csb	2011–14	936
Velas**	São Jorge	38° 39′ 51.08″	28° 10′ 14.94″	99	4.1	Csa	2012–13	283
Santa Cruz da Graciosa**	Graciosa	39° 5′ 29.76″	28° 1′ 32.52″	30	2.0	Csa	2013–17	1307
Angra do Heroísmo	Terceira	38° 39′ 31.68″	27° 13′ 22.8″	74	2.9	Csb	2003–13	2607
Granja	Terceira	38° 41′ 45.88″	27° 10′ 14.51″	370	4.8	Csb	1996–98 and 2002–11	4619
Lajes**	Terceira	38° 45′ 21.68″	27° 04′ 59.86″	54	3.3	Csb	2000–14	5478
Ribeirinha	Terceira	38° 40′ 24.16″	27° 10′ 43.27″	390	5.1	Csb	1996–97 and 2002–10	3780
Santa Bárbara	Terceira	38° 43′ 38.85″	27° 19′ 47.81″	800	7.1	Csb	2002–11	3594
Chã Macela	S. Miguel	37° 45′ 52.49″	25° 31′ 55.24″	310	3.4	Csa	2010–14	1095
Lagoa do Fogo	S. Miguel	37° 45′ 50.03″	25° 29′ 43.98″	890	7.5	Csb	2007–12	1612
Furnas	S. Miguel	37° 45′ 42.56″	25° 19′ 44.05″	289	5.9	Csb	2006–14	3922
Ponta Delgada**	S. Miguel	37° 44′ 38.76″	25° 42′ 28.80″	71	2.9	Csa	2002–14	3348
Santana	S. Miguel	37° 48′ 17.62″	25° 33′ 42.65″	70	4.0	Csa	2000–12	3376
Sete Cidades	S. Miguel	37° 52′ 4.60″	25° 46′ 19.37″	260	6.1	Csa	2002–09	2167
Tronqueira	S. Miguel	37° 46′ 9.91″	25° 11′ 51.14″	500	5.9	Csb	2010–14	1332
Maia	Santa Maria	36° 56′ 25.38″	25° 1′ 3.18″	200	3.7	Csa	2011–17	2249
Praia Formosa	Santa Maria	36° 57′ 9.79″	25° 5′ 35.81″	180	2.8	Csa	2011–14	1183
Fontinhas	Santa Maria	36° 57′ 45.52″	25° 4′ 34.27″	420	3.6	Csa	2010–14	1282

*Potential evapotranspiration (PET) computed with Thornthwaite equation

**Weather stations located at airports

other side of the island, the descending air mass decreases humidity and increases temperature as for the foehn effect. The model was validated to the Terceira Island (Azevedo et al. 1999) and tested in the S. Miguel Island (Santos et al. 2004; Miranda et al. 2006). Model results consist of the spatial distribution of all simulated climatic variables on the island territory, thus CIELO was applied to all islands to support ecological studies, e.g., vegetation distribution (Elias et al. 2016). Because solar radiation is not simulated, it is important that, using the findings of the current study, both R_s and ET_o could be estimated using the reduced data sets created by CIELO. Using CIELO in climate change studies (Santos et al. 2004; Miranda et al. 2006) it was concluded that main impacts of global warming will be on the intra-annual precipitation distribution, with wetter winters and other seasons becoming drier, which will impact the Azores water resources and their availability for vegetation and crops.

2.2 Reference evapotranspiration computations

The PM-ET_o Eq. (1) corresponds to definition of the grass reference ET_o. Computations require the adoption of the standard methods proposed by Allen et al. (1998) for computing the various parameters of the equation. As analyzed by Nandagiri and Kovoor (2005), results of the PM-ET_o Eq. (1) using different computational approaches for the respective parameters differ from those obtained when standard computation procedures are adopted. Equation (1) is used herein as standard for assessing the performance of the alternative approaches to compute ET_o with reduced data sets.

Following Allen et al. (1998), the standard computation of the vapor pressure deficit (VPD, kPa), which is the difference between the saturation vapor pressure (e_s, kPa) and the actual vapor pressure (e_a, kPa), was performed with e_s given as

$$e_s = \frac{e^o(T_{max}) + e^o(T_{min})}{2} \tag{7}$$

where $e^o(T_{max})$ and $e^o(T_{min})$ are the saturation vapor pressure at respectively the maximum and minimum daily temperatures (kPa), and e_a was computed as

$$e_a = \frac{e^o(T_{min})\frac{RH_{max}}{100} + e^o(T_{max})\frac{RH_{min}}{100}}{2} \tag{8}$$

Table 2 Weather stations data available and respective institutions in charge

Weather stations	Island	Observations of RH (%)	Observed R_s (W m^{-2})	Observed sunshine hours, n (h)	Anemometer height (m)	Institutions in charge
Santa Cruz das Flores	Flores	Average	Yes		10	IPMA
Horta	Faial	Average	Yes		10	IPMA
São Caetano	Pico	Average	Yes		3	SRAM-Açores
Velas	São Jorge	Average	Yes		3	SRAM-Açores
Santa Cruz da Graciosa	Graciosa	Average	Yes		10	ARM
Angra do Heroísmo	Terceira	Average	Yes		10	IPMA
Granja	Terceira	Maximum, minimum	Yes		2	UAçores
Lajes	Terceira	Average		Yes	2	IPMA
Ribeirinha	Terceira	Maximum, minimum	Yes		2	UAçores
Santa Bárbara	Terceira	Maximum, minimum	Yes		2	UAçores
ChãMacela	S. Miguel	Average	Yes		3	SRAM-Açores
Lagoado Fogo	S. Miguel	Average	Yes		3	SRAM-Açores
Furnas	S. Miguel	Average	Yes		2	SRAM-Açores
Ponta Delgada	S. Miguel	Average	Yes		10	IPMA
Santana	S. Miguel	Average	Yes		3	SRAM-Açores
Sete Cidades	S. Miguel	Average	Yes		3	SRAM-Açores
Tronqueira	S. Miguel	Average	Yes		3	SRAM-Açores
Maia	Santa Maria	Maximum, minimum	Yes		3	SRAM-Açores
Praia Formosa	Santa Maria	Average	Yes		3	SRAM-Açores
Fontinhas	Santa Maria	Average	Yes		3	SRAM-Açores

RH, relative humidity; R_s, Solar radiation; IPMA, Instituto Português do Mar e da Atmosfera; ARM, Atmospheric Radiation Measurement, US Department of Energy; SRAM-Açores, Secretaria Regional do Ambiente e do Mar, Açores; UAçores, Universidade dos Açores

in locations where observations of maximum and minimum relative humidity, RH_{max}, and RH_{min} (%) were available (Table 2), or as

$$e_a = \frac{RH_{mean}}{100}\left[\frac{e^o(T_{max}) + e^o(T_{min})}{2}\right] \tag{9}$$

when only the mean daily RH (RH_{mean}, %) was available (Table 2).When RH data are missing, given the humid environments of the islands, the approach described with Eq. 5 was adopted. Its results were compared with those obtained with the standard Eqs. 8 or 9 depending upon data observed. T_{dew} was firstly derived from observed e_a as

$$T_{dew} = \frac{116.91 + 237.3\ln(e_a)}{16.78 - \ln(e_a)} \tag{10}$$

and the derived T_{dew} was compared with T_{mean} to derive the adjustment factors a_d to be used with Eq. 4. To improve accuracy of ET_o estimations, the values of a_d were derived for the semesters October–March (Winter) and April–September (Summer) seasons, which relate with the intra-annual variation of the precipitation and temperature with winter being the period having most of precipitation and lower temperature (Fig. 2).

Solar radiation, R_s, was observed in most weather stations (Table 2) and sunshine duration was observed only in Lajes. For the latter, R_s was calculated as (Allen et al. 1998):

$$R_s = \left(a_s + b_s\frac{n}{N}\right)R_a \tag{11}$$

where R_s is shortwave solar radiation (MJ m^{-2} day^{-1}), n is the actual sunshine duration (hour), N is maximum possible sunshine duration or daylight hours (hour), n/N is relative sunshine duration (−), R_a is extraterrestrial radiation (MJ m^{-2} day^{-1}), a_s is the fraction of extraterrestrial radiation reaching the earth on overcast days ($n = 0$), and $a_s + b_s$ is the fraction of extraterrestrial radiation reaching the earth on clear-sky days ($n = N$). The default parameters $a_s = 0.25$ and $b_s = 0.50$ were used as recommended by Allen et al. (1998).

The calibration of the radiation adjustment factor k_{Rs} (Eq. 6) was performed using a trial and error procedure through an iteratively adjusting the k_{Rs} value when comparing R_s computed with Eq. 6 with R_s observations. For every location, this procedure was applied seasonally as for the calibration of the factor a_d as referred above.

To adjust wind speed data obtained from instruments placed above the standard height of 2 m, a logarithmic wind speed profile (Allen et al. 1998) was used:

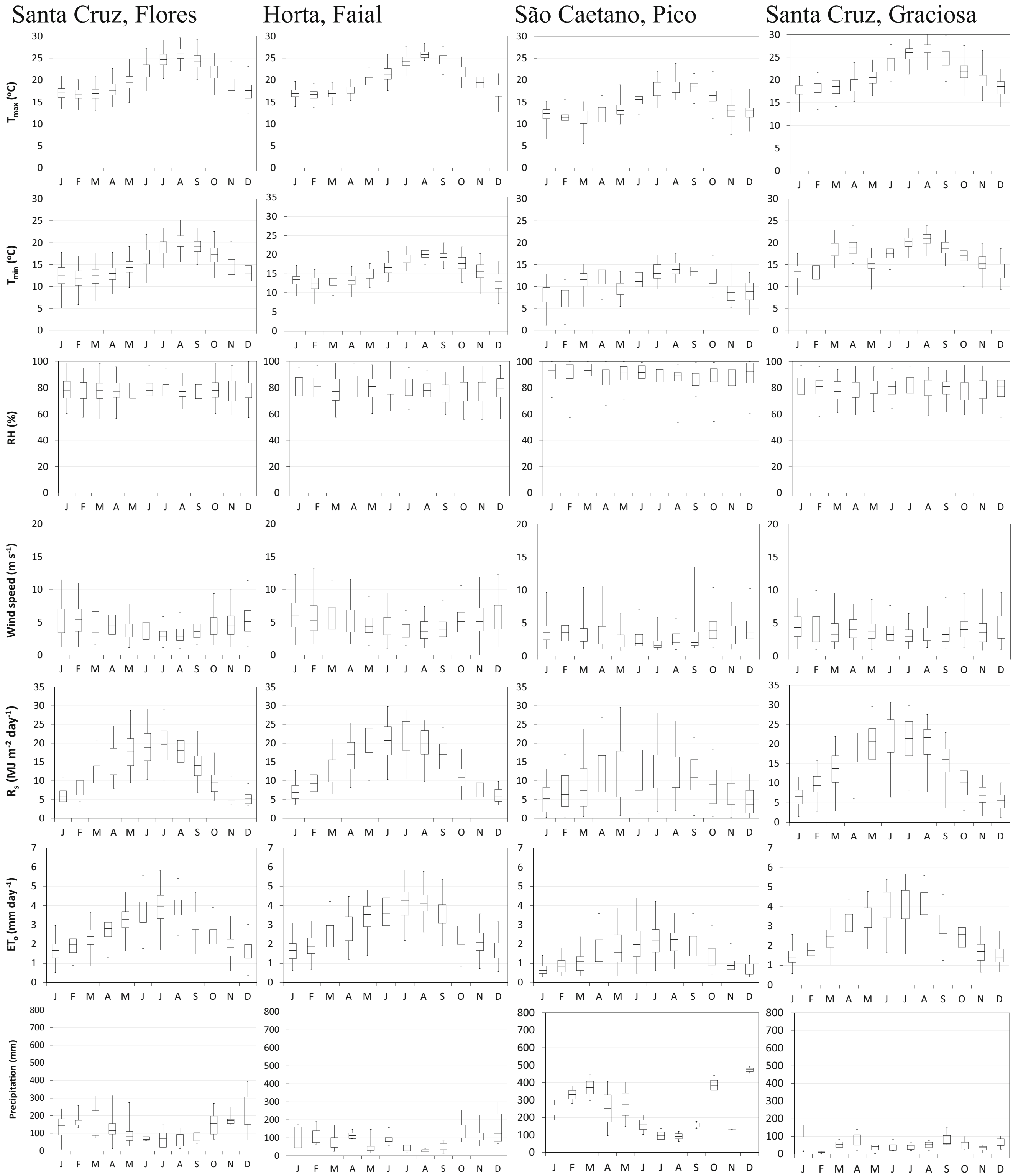

Fig. 2 Box-and-whisker plot showing the lower quartile (Q1), the median (Q2), and the upper quartile (Q3) as well as the minimum and maximum values relative to the main weather variables for selected locations. Data relative to each weather station refer to the periods of observation indicated in Table 1

$$u_2 = u_z \frac{4.87}{\ln(67.8\, z - 5.42)} \tag{12}$$

where u_2 is wind speed at 2 m above ground surface (m s^{-1}), u_z is the measured wind speed at z m height (m s^{-1}), and z is

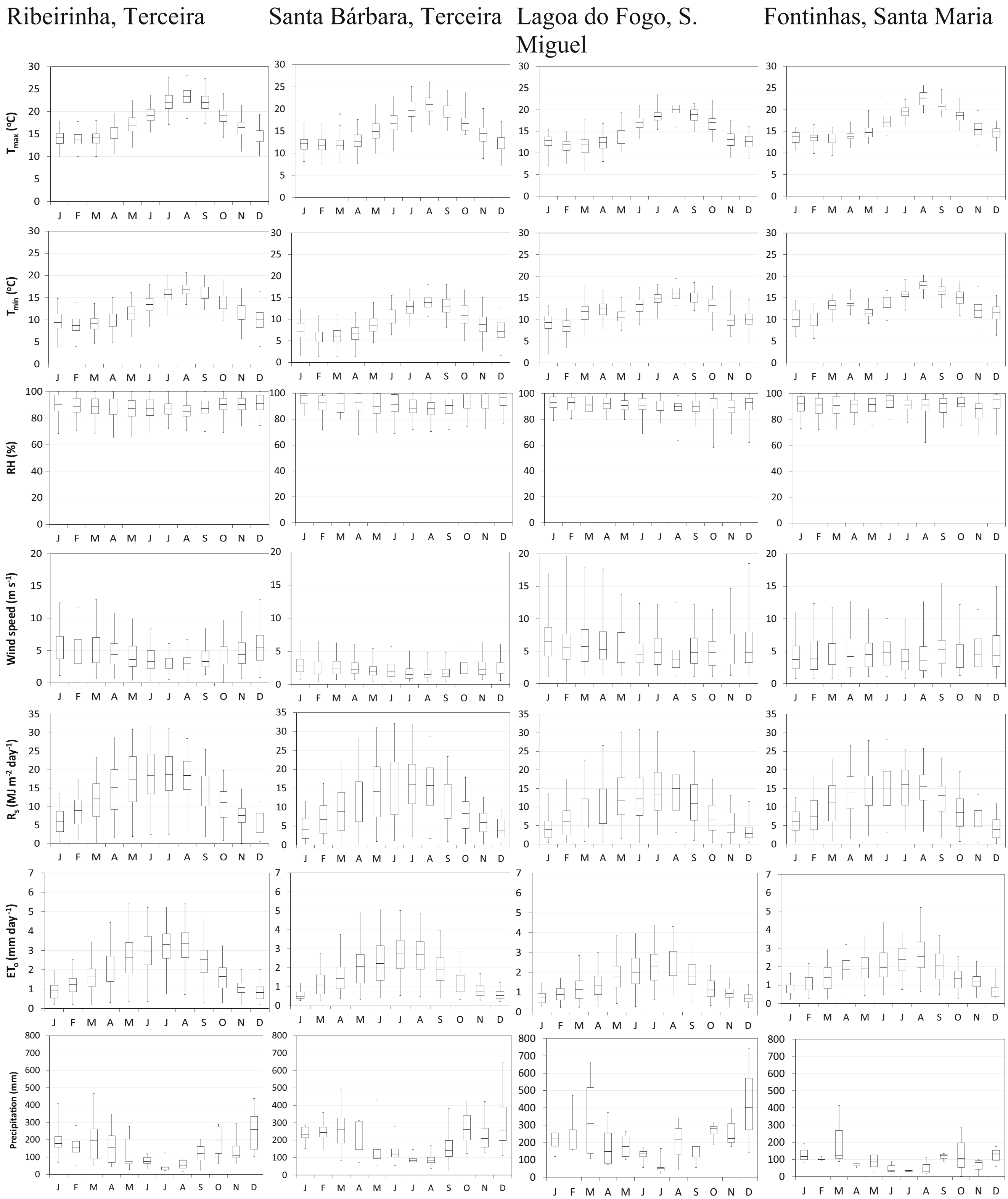

Fig. 2 (continued)

height of measurement above ground surface (m).When wind speed data are missing, u_2 was estimated through the local average of wind speed and adopting the default $u_{2def} = 2$ m s^{-1} or 3 m s^{-1}, the latter in case of weather measurements over grass in airports, where wind speed is typically high (Fig. 2).

For all three cases described above, ET_o computed with the estimated missing variables e_a using T_{dew} estimates, $ET_{o\ Tdew}$, R_s using the temperature difference method, $ET_{o\ TD}$, and u_2 with the yearly average wind speed or the default u_2, respectively u_{2avg} and u_{2def}, thus the $ET_{o\ uavg}$ and $ET_{o\ udef}$ were assessed for accuracy against the PM-ET_o.

2.3 Accuracy indicators

The accuracy of ET_o computations using alternative approaches for estimating e_a, R_s, and u_2, was assessed by comparing their results with those of the PM-ET_o equation using full data sets. In addition, the estimators of e_a using estimated T_{dew} values ($e_{a\ Tdew}$) and of R_s using the calibrated k_{Rs} and the temperature difference equation ($R_{s\ TD}$) were compared with e_a and R_s obtained from observations. Following previous applications to ET_o studies, accuracy was measured with several statistical indicators as described by Martins et al. (2017):

1. The regression coefficient (b_0) of the regression forced to the origin (FTO, Eisenhauer, 2003) between daily PM-ET_o computed with observed data, O_i (x), and daily ET_o computed with estimated variables, P_i (y). A value of b_0 = 1.0 indicates that the fitted line is y = x, so that O_i and P_i are similar. A value of $b_0 > 1.0$ suggests over-estimation and $b_0 < 1.0$ under-estimation.
2. The determination coefficient (R^2) of the ordinary least squares (OLS) regression between O_i and P_i. The higher the value of R^2, the more variation of the PM-ET_o values is explained by the simplified computation approach; however, a high value of R^2 is, in itself, insufficient to state that there is good overall agreement between observed and estimated values.
3. The root mean square error (RMSE, mm day^{-1}) that measures overall discrepancies between observed and estimated values, hence the smaller it is the better is the accuracy.
4. The percent bias (PBIAS, %), which is simply a normalized difference between the means of both sets O_i and P_i, and as a bias indicator measures the average tendency of the simulated data to be larger or smaller than their corresponding observations.
5. The Nash and Sutcliffe (1970) modeling efficiency (EF, non-dimensional) that provides an indication of the relative magnitude of the mean square error (MSE = RMSE^2) relative to the observed data variance (Legates and McCabe 1999), i.e., compares "noise" with "information" (Moriasi et al. 2007). EF is estimated as

$$EF = 1-\frac{\sum_{i=1}^{M}(O_i-P_i)^2}{\sum_{i=1}^{M}\left(O_i-\overline{O}\right)^2} \tag{13}$$

The maximum value EF = 1.0 can only be achieved if there is a perfect match between all observed (O_i) and predicted (P_i) values, thus with RMSE = 0, R^2 = 1.0, and b_0 = 1. The closer the values of EF are to 1.0, the more "noise" is negligible relative to the "information," implying that alternative-based values of ET_o are good estimators of PM-ET_o values. Negative values of EF indicate that MSE is larger than the observed data variance meaning that it would be better to use the mean of observed values rather than the predicted values P_i as Legates and McCabe (1999) correctly stressed.

As shown by Martins et al. (2017), the joint assessment of this set of indicators provides a good evaluation of the quality of prediction of weather variables and ET_o when using the referred approaches to estimate the missing weather variables. Scatter plots and the regression lines relating O_i and P_i were also analyzed for every variable set. Eight examples of graphical results are presented to support discussion. They were selected in seven islands and covering various environmental conditions.

3 Results and discussion

3.1 Estimation of ET_o in the absence of RH observations

In the absence of RH data, the actual vapor pressure (e_a) may be estimated using T_{min} as estimator of T_{dew} (Eq. 2). However, in humid climates $T_{dew} > T_{min}$ and is better related with T_{mean}, i.e., $T_{dew} = T_{mean} - a_d$ (Eq. 4), which requires appropriate calibration of a_d. Computations using $T_{dew} = T_{min}$ revealed acceptable accuracy but a tendency for under-estimation of ET_o in most locations (results not shown). Differently, using a calibrated adjustment factor a_d, the regression between ($T_{mean} - a_d$) and T_{dew} has shown regression coefficients ranging from 0.96 to 1.00, coefficients of determination $R^2 > 0.81$, small RMSE < 1.69 °C, PBIAS in the interval − 1.2 to 2.4%, and EF > 0.81. Examples of scatter plots representing that regression for the eight selected locations (Fig. 3) support the assumption that a linear relation describes well the strong relationship between ($T_{mean} - a_d$) and T_{dew}. It may be observed that, however, ($T_{mean} - a_d$) tends to over-estimate T_{dew} when air temperature is low but not for medium and higher temperatures.

A regression analysis was also used to compare e_a computed from the estimated T_{dew}, $e_{a\ Tdew}$, with e_a computed from RH observations. Results for the referred eight selected locations (Fig. 4) show that respective results are very close statistically, with b_0 ranging from 0.95 to 0.98, and $R^2 > 0.82$, nevertheless with a slight under-estimation for the high e_a values and over-estimation for the low ones. In addition, results for all locations present small RMSE < 0.18 kPa, PBIAS in the interval

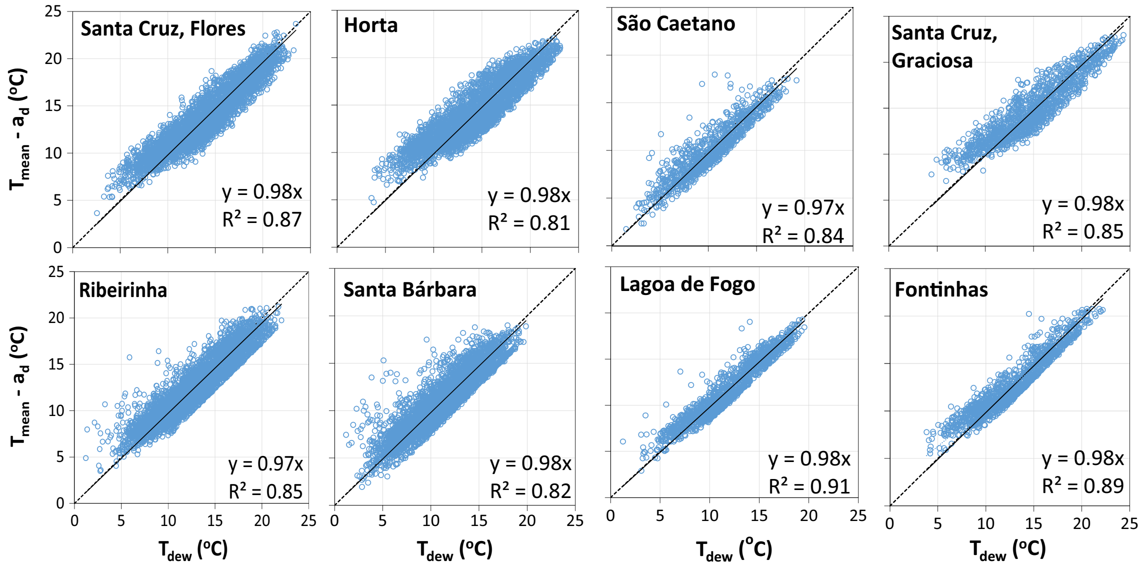

Fig. 3 Relationships between ($T_{mean} - a_d$) and T_{dew} for eight selected locations in various islands and having different environments. Included the FTO regression equation and the OLS determination coefficient R^2; also depicted the 1:1 line (- - -)

− 0.6 to 3.5%, and EF > 0.82 indicate that the estimation of e_a from $T_{dew} = T_{mean} - a_d$ is a very good approach for islands' humid environments.

Testing T_{dew} estimations were performed by comparing $ET_{o\ Tdew}$, computed with $e_{a\ Tdew}$, with PM-ET_o computed with observed full data sets. Results in Table 3 show that a_d values ranged from 1.5 to 4 °C and that for most locations, a_d values were the same for both seasons; when seasonal differences were observed, the corresponding a_d differences were small (0.5 °C),with the lower values for the more humid winter. Higher a_d values were more frequent for western locations, where rainfall and climate humidity are higher, and lower values are often for locations of medium to high elevation, also exposed to the wind. However, data available are insufficient to derive a proper relationship between a_d values and selected characteristic of the stations.

The accuracy of daily $ET_{o\ Tdew}$ estimations are presented in Table 3 and Fig. 5. Results show to be very good considering the goodness of results for estimating T_{dew} as well as $e_{a\ Tdew}$ (Figs. 3 and 4). No tendency for under- or over-estimation were observed in 70% of stations, where b_0 values are in the interval 0.99–1.01 and all b_0 values range from 0.98 to 1.02. Consequently, PBIAS values are more often within the interval of − 5 to 5%, with only three locations presenting an under-estimation bias ranging between − 7 and − 5%. Large R^2 values ($R^2 > 0.90$) are observed in 55% of locations and

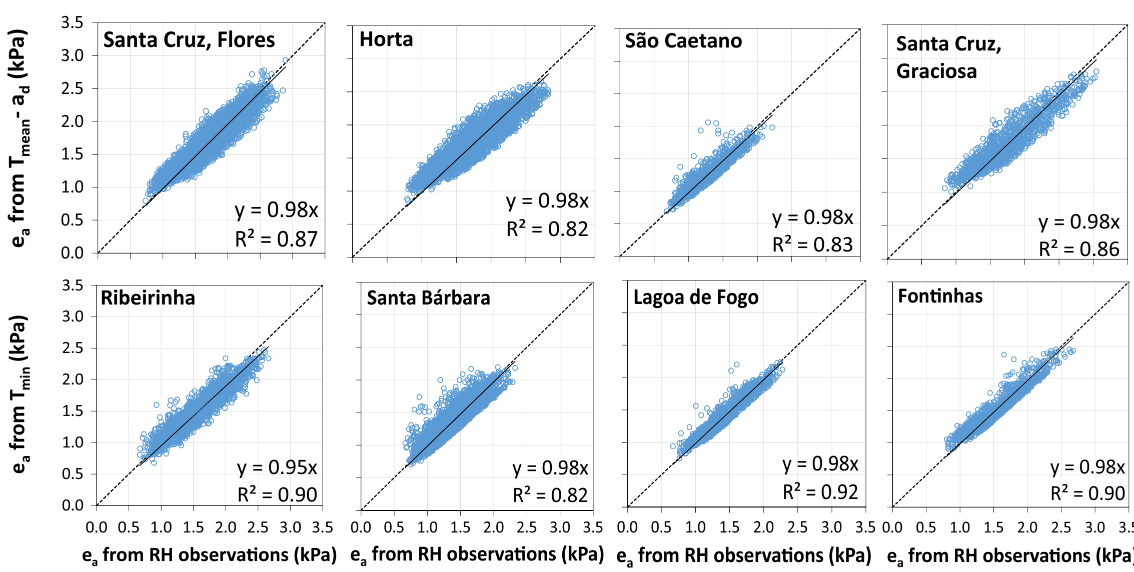

Fig. 4 Comparing $e_{a\ Tdew}$, computed from ($T_{mean} - a_d$) and e_a, computed with observed data, for eight selected locations in various islands and having different environments. Included the FTO regression equation and the OLS determination coefficient R^2; also depicted the 1:1 line (- - -)

Table 3 Calibrated values for the temperature adjusting factor a_d and accuracy indicators of ET_o estimations in the absence of RH data using $T_{dew} = T_{mean} - a_d$ (Eq. 4)

Weather station	$T_{dew} = T_{mean} - a_d$	Statistical indicators				
		b_0	R^2	RMSE (mm day^{-1})	PBIAS (%)	EF
Santa Cruz das Flores	$a_d = 4$	1.00	0.89	0.37	−2.1	0.89
Horta	$a_d = 3.5$	0.98	0.86	0.46	−2.6	0.86
São Caetano	Winter $a_d = 1.5$ Summer $a_d = 2$	0.98	0.90	0.29	−4.0	0.89
Velas	$a_d = 3.5$	1.00	0.83	0.40	−3.5	0.82
Santa Cruz da Graciosa	$a_d = 3.5$	0.99	0.85	0.45	−1.6	0.85
Angra Heroísmo	$a_d = 3.5$	1.00	0.92	0.36	−2.3	0.91
Granja	$a_d = 2$	0.99	0.94	0.32	−4.7	0.93
Lajes	Winter $a_d = 3.5$ Summer $a_d = 4$	1.01	0.79	0.53	−4.2	0.78
Ribeirinha	Winter $a_d = 2$ Summer $a_d = 2.5$	0.99	0.85	0.48	−6.9	0.83
Santa Bárbara	Winter $a_d = 1$ Summer $a_d = 2$	0.99	0.95	0.25	−4.3	0.94
Chã Macela	$a_d = 2.5$	1.01	0.97	0.17	−3.3	0.96
Lagoa Fogo	$a_d = 1.5$	1.00	0.96	0.18	−3.3	0.96
Furnas	Winter $a_d = 2.5$ Summer $a_d = 3$	1.01	0.93	0.28	−3.5	0.92
Ponta Delgada	$a_d = 3.5$	0.98	0.86	0.46	−1.4	0.86
Santana	$a_d = 3$	0.98	0.96	0.21	0.4	0.95
Sete Cidades	$a_d = 2.5$	1.02	0.92	0.25	−5.7	0.90
Tronqueira	Winter $a_d = 1.5$ Summer $a_d = 2$	1.00	0.98	0.15	−2.8	0.98
Maia	Winter $a_d = 3$ Summer $a_d = 3.5$	0.99	0.94	0.28	−0.8	0.94
Praia Formosa	$a_d = 3$	0.99	0.84	0.40	−3.0	0.83
Fontinhas	$a_d = 1.5$	0.98	0.83	0.38	−5.2	0.81

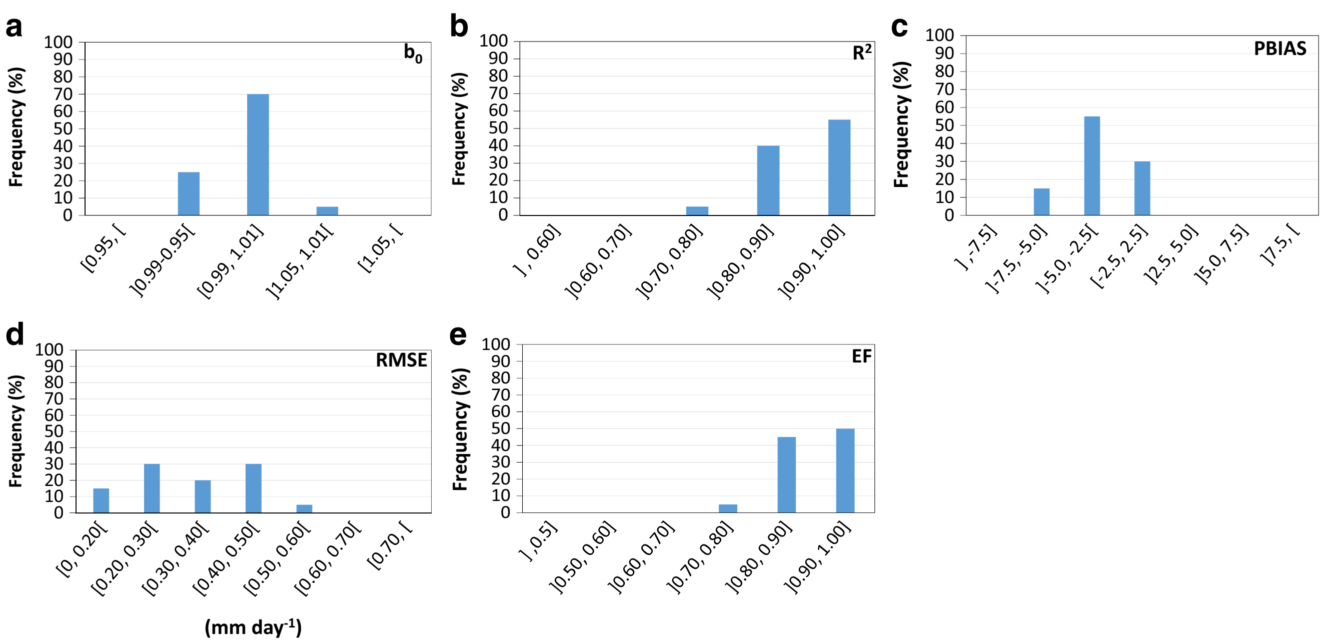

Fig. 5 Frequency distribution of the accuracy indicators relative to the computation of ET_{oTdew} when the actual vapor pressure computed from $T_{dew} = T_{mean} - a_d$

$R^2 > 0.80$ was for 95% of cases, thus meaning that $ET_{o\ Tdew}$ well explains most of the variation of the PM-ET_o results computed with full data sets. In addition, RMSE are generally small, ranging from 0.15 to 0.53 mm day^{-1} with 65% of stations with RMSE < 0.40 mm day^{-1}. Such small RMSE values indicate that the discrepancies between $ET_{o\ Tdew}$ and PM-ET_o are small. The EF values were very high, with EF > 0.90 in 50% of the weather stations and EF > 0.80 in 95% of the locations, thus meaning that MSE are much lower than the variance of PM-ET_o results. Hence, using $T_{dew} = T_{mean} - a_d$ with a_d values listed in Table 2 to estimate actual vapor pressure is appropriate to overcome problems due to missing RH data or when they have no adequate quality.

The scatter plots in Fig. 6 not only highlight the very good relationships between $ET_{o\ Tdew}$ and PM-ET_o but also allow to perceive which data shows slight inequalities between both computational approaches: slight over-estimations occur for the smaller values of ET_o, and negligible under-estimations occur for the high ET_o values. This behavior is well explained by the relationships between $e_{a\ Tdew}$ and e_a analyzed before (Fig. 4).

Few studies report on the accuracy of ET_o estimations in the absence of RH, and less are available for humid climates but different of that of Azores islands. Better indicators were reported using $T_{dew} = T_{min}$ for sub-humid climates, e.g., by Liu and Pereira (2001) and Pereira et al. (2003) for North China Plain, and Popova et al. (2006) for Thrace, Bulgaria, as well as by Trajkovic and Kolakovic (2009) for Serbia, and Sentelhas et al. (2010) and Aladenola and Madramootoo (2014) for humid sites of Canada. However, it is not possible to properly compare with climatic conditions as those in Azores where intra-annual and monthly variability of air masses flowing over the islands causes a great variability of climate variables (Fig. 2).

3.2 Estimation of ET_o when solar radiation data is not available

As previously stated, when R_s observations are not available, estimations of ET_o may be performed using the Hargreaves temperature difference equation (Eq. 6) with a locally calibrated radiation adjustment coefficient k_{Rs}, thus resulting in the $ET_{o\ TD}$ equation.

Results in Table 3 show that k_{Rs} values ranged from 0.14 to 0.25 °C$^{-0.5}$with lower k_{Rs} values in locations at medium and high altitude. Because the coefficient k_{Rs} is supposed to reflect the volumetric heat capacity of the atmosphere (Allen 1997), therefore, lower k_{Rs} values are to be expected with elevation. As discussed by Turbet et al. (2017), the volumetric heat capacity of the atmosphere increases linearly with the volumetric mass density and thus with the atmospheric pressure. For that reason, Annandale et al. (2002) developed a decreasing relationship between k_{Rs} and elevation to account for the effects of reduced atmospheric thickness. The occurrence of winds transporting humid air masses may also contribute to decrease the heat capacity of the atmosphere, as already described with model CIELO (Azevedo et al. 1999). Nevertheless, the sample size is small and does not allow to develop a statistically significant relationship between k_{Rs} and relevant station characteristics.

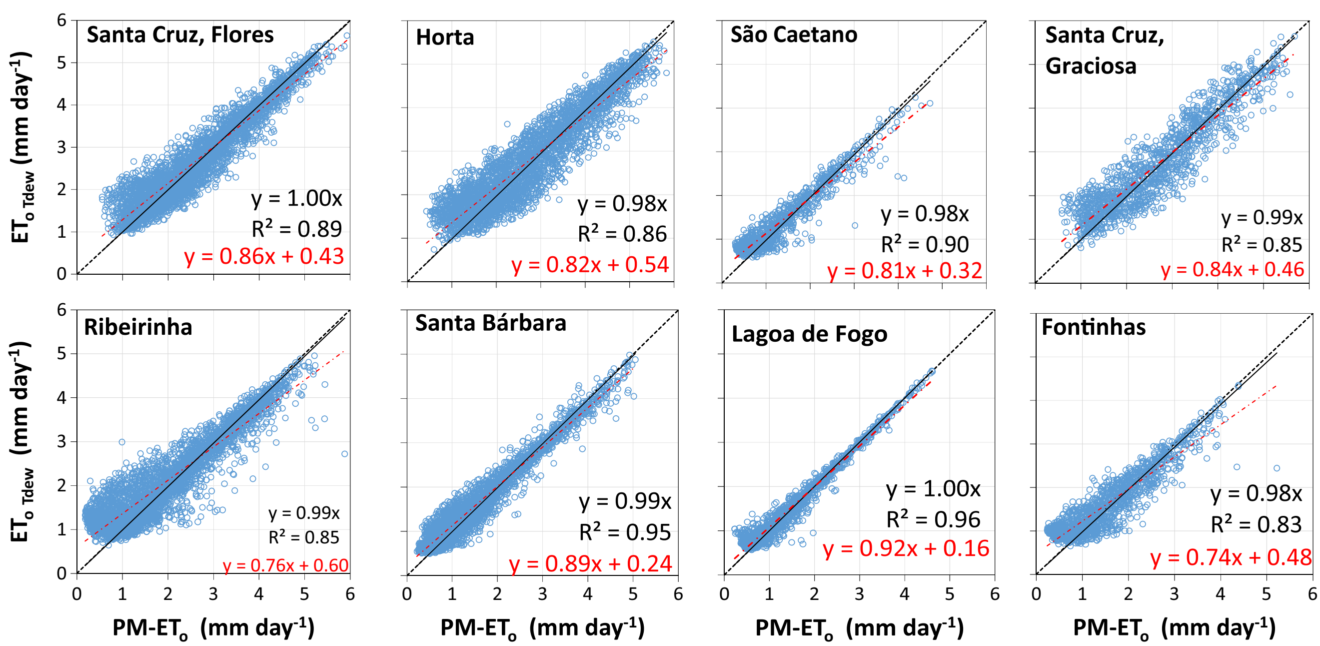

Fig. 6 Comparing $ET_{o\,Tdew}$, computed with the estimator $e_{a\ Tdew}$ estimated from ($T_{dew} = T_{mean} - a_d$) and PM-ET_o computed with full data sets, for eight selected locations in various islands and having different environments. Included the FTO and the OLS regression equations and the OLS determination coefficient R^2; the 1:1 line (- - -) and the OLS line (- . - . -) are depicted

Results of the calibration of k_{Rs} show that in 45% of locations, different k_{Rs} values where found for the winter and summer seasons, with lower k_{Rs} in winter (Table 4). Seasonal differences are likely due to the fact that for those locations, the volumetric heat capacity of the atmosphere is higher in summer since the frequent cloud cover may limit it during winter. Computations for a period covering both seasons were performed assigning different k_{Rs} to the winter and summer periods.

Results relative to R_s estimations with Eq. 6 resulted in a trend for under-estimation with RMSE ranging 3.5 to 5.5 MJ m^{-2} day^{-1}. Selected examples of regression of $R_{s\ TD}$ with observed R_s are presented in Fig. 7. No study is available for conditions similar to those in Azores and few applications for humid to sub-humid conditions are available in literature. Results of our study are in the range of those reported by Aladenola and Madramootoo (2014) and Kwon and Choi (2011); better results were reported by Almorox et al. (2016) and Lyra et al. (2016) when calibrating k_{Rs}.

Accuracy indicators (Table 4, Fig. 8) show that in 50% of locations, b_0 ranged from 0.99 to 1.00 and that in the other 50% of cases, b_0 was in the range of 0.96 to 0.98, hence meaning that $ET_{o\ TD}$ and PM-ET_o are statistically very close, however with $ET_{o\ TD}$ slightly underestimating PM-ET_o. It resulted that the PBIAS values are generally negative but small, with PBIAS not exceeding − 7%, generally in the interval − 5 to 5%. Higher PBIAS were observed in locations with medium to high altitude, which may relate with the high cloudiness of stations and combined impacts of winds carrying humid air masses. R^2 values were generally close to 1.0, with $R^2 > 0.80$ in 80% of cases, thus meaning that most of variation of PM-ET_o can be explained with the $ET_{o\ TD}$ approach. RMSE are quite small with 95% of the locations having RMSE < 0.50 mm day^{-1}. The EF values are high, ranging 0.70 to 0.91, with EF > 0.80 in 80% of locations, indicating that the mean square error is much lower than the variance of ET_o computed with observed full data sets.

The scatter plots in Fig. 9 highlight the very good relationships between $ET_{o\ TD}$ and PM-ET_o, despite the less good accuracy in $R_{s\ TD}$ estimations (Fig. 7) but also allow perceiving that a slight over-estimation occur for the higher values of ET_o, and in most locations a negligible under-estimations

Table 4 Accuracy of ET_o estimation in the absence of R_s observations using temperature difference (Eq. 6) and the calibrated k_{Rs} value

Weather stations	$R_s = k_{R_s}(T_{max} - T_{min})^{0.5} R_a$	b_0	R^2	RMSE (mm day^{-1})	PBIAS (%)	EF
Santa Cruz das Flores	Winter $k_{Rs} = 0.19$ °C$^{-0.5}$ Summer $k_{Rs} = 0.21$ °C$^{-0.5}$	0.98	0.86	0.41	− 0.7	0.86
Horta	$k_{Rs} = 0.25$ °C$^{-0.5}$	0.99	0.89	0.42	− 3.6	0.88
São Caetano	$k_{Rs} = 0.18$ °C$^{-0.5}$	0.98	0.81	0.38	− 6.0	0.80
Velas	$k_{Rs} = 0.20$ °C$^{-0.5}$	1.00	0.87	0.35	− 2.8	0.86
Santa Cruz da Graciosa	Winter $k_{Rs} = 0.19$ °C$^{-0.5}$ Summer $k_{Rs} = 0.22$ °C$^{-0.5}$	0.98	0.91	0.36	0.4	0.91
Angra Heroísmo	Winter $k_{Rs} = 0.22$ °C$^{-0.5}$ Summer $k_{Rs} = 0.23$ °C$^{-0.5}$	0.98	0.86	0.45	− 1.2	0.86
Granja	$k_{Rs} = 0.17$ °C$^{-0.5}$	0.98	0.87	0.42	− 1.8	0.87
Lajes	Winter $k_{Rs} = 0.20$ °C$^{-0.5}$ Summer $k_{Rs} = 0.21$ °C$^{-0.5}$	0.99	0.90	0.36	− 0.8	0.90
Ribeirinha	$k_{Rs} = 0.20$ °C$^{-0.5}$	1.00	0.90	0.38	− 2.9	0.89
Santa Bárbara	Winter $k_{Rs} = 0.15$ °C$^{-0.5}$ Summer $k_{Rs} = 0.17$ °C$^{-0.5}$	0.99	0.84	0.44	− 6.9	0.83
Chã Macela	Winter $k_{Rs} = 0.15$ °C$^{-0.5}$ Summer $k_{Rs} = 0.17$ °C$^{-0.5}$	0.99	0.75	0.46	− 4.0	0.74
Lagoa do Fogo	Winter $k_{Rs} = 0.20$ °C$^{-0.5}$ Summer $k_{Rs} = 0.21$ °C$^{-0.5}$	0.98	0.71	0.41	− 7.0	0.70
Furnas	Winter $k_{Rs} = 0.19$ °C$^{-0.5}$ Summer $k_{Rs} = 0.20$ °C$^{-0.5}$	1.00	0.79	0.47	− 4.4	0.78
Ponta Delgada	$k_{Rs} = 0.24$ °C$^{-0.5}$	0.98	0.89	0.42	− 0.3	0.88
Santana	$k_{Rs} = 0.19$ °C$^{-0.5}$	0.99	0.81	0.44	− 3.1	0.81
Sete Cidades	Winter $k_{Rs} = 0.14$ °C$^{-0.5}$ Summer $k_{Rs} = 0.15$ °C$^{-0.5}$	0.96	0.82	0.33	− 0.7	0.82
Tronqueira	$k_{Rs} = 0.16$ °C$^{-0.5}$	0.97	0.83	0.39	− 3.6	0.83
Maia	$k_{Rs} = 0.21$ °C$^{-0.5}$	0.98	0.77	0.58	− 3.2	0.76
Praia Formosa	$k_{Rs} = 0.20$ °C$^{-0.5}$	1.00	0.86	0.36	− 2.5	0.86
Fontinhas	$k_{Rs} = 0.22$ °C$^{-0.5}$	0.99	0.83	0.37	− 4.1	0.82

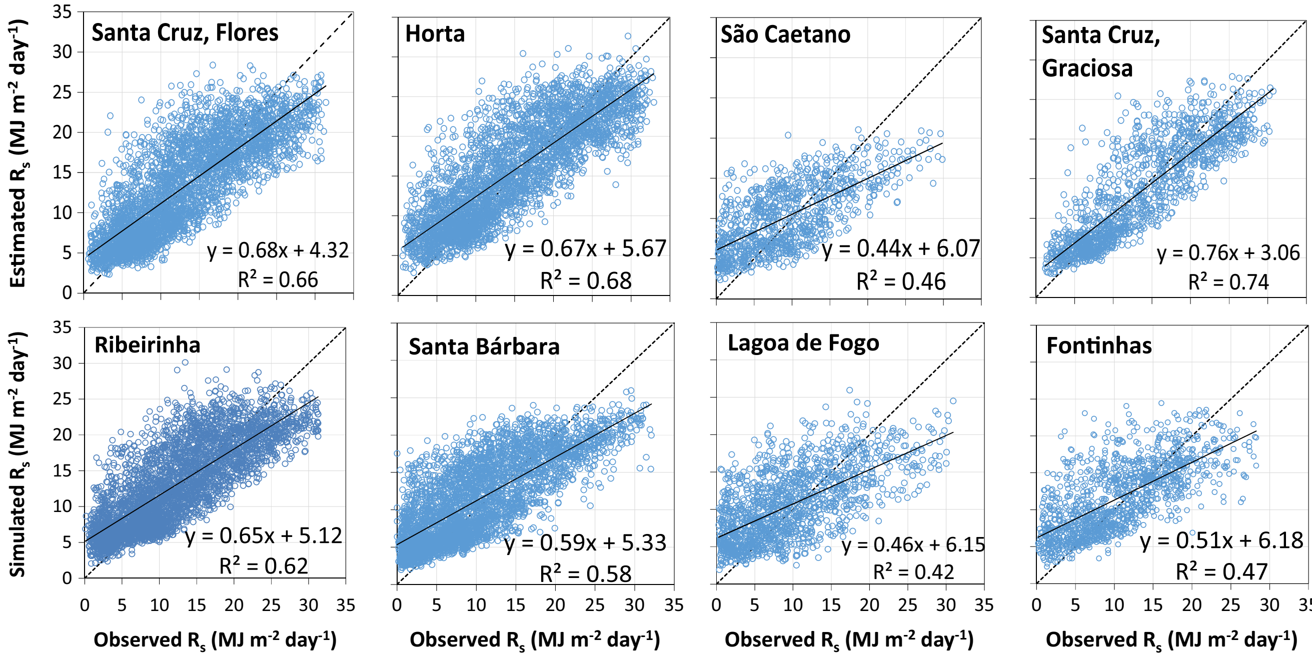

Fig. 7 Selected examples comparing the estimated solar radiation $R_{s\ TD}$ computed with the temperature difference and calibrated k_{Rs} (Eq. 6) with and R_s from observations. Included the OLS regression equation and the respective R^2; also depicted the 1:1 line (- - -) and the OLS line (—)

occur for the low ET_o values. This behavior is well explained by the relationships between $R_{s\ TD}$ and observed R_s as previously analyzed (Fig. 7).

The results of the present study are difficult to compare with other studies where only R_s estimates from the temperature difference is used to compute ET_o in humid island environments. Examples from humid climates, but different from the Azores environment, are available. Smaller RMSE were observed by Popova et al. (2006) and Aladenola and Madramootoo (2014) using a constant value for k_{Rs}. Pereira et al. (2003), Sentelhas et al. (2010), Kwon and Choi (2011), and Lyra et al. (2016) reported higher RMSE values than the current study. No other studies were found for islands with humid climate.

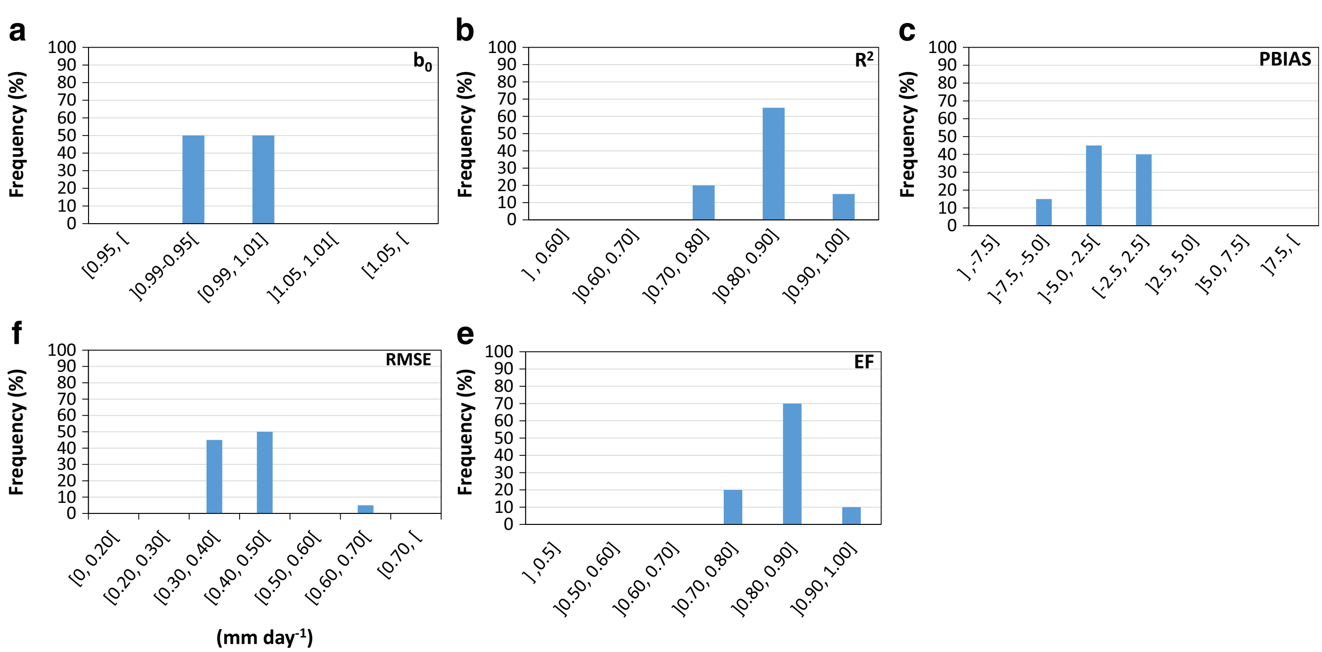

Fig. 8 Frequency distribution of the accuracy indicators relative to the computation of ET_o using $R_{s\ TD}$ derived from the temperature difference and a calibrated radiation adjustment coefficient k_{Rs}

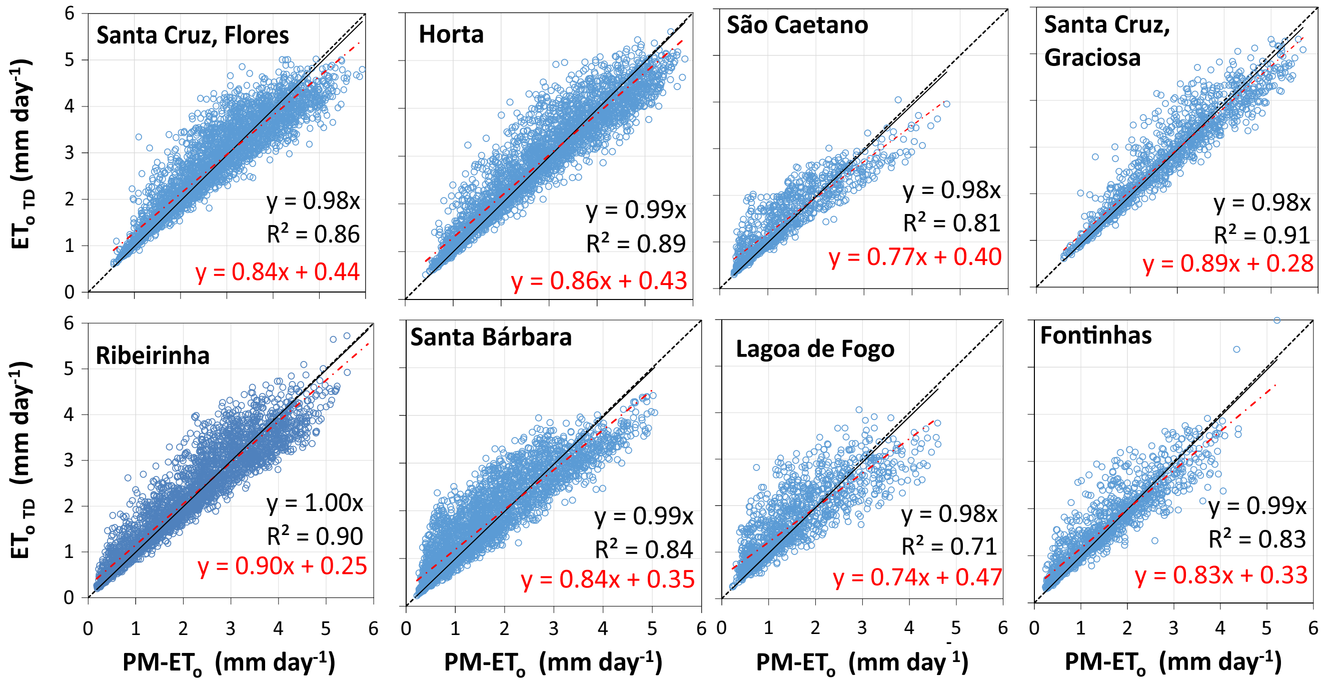

Fig. 9 Selected examples of comparing $ET_{o\ TD}$, computed with the temperature difference equation and adopting calibrated k_{Rs}. The FTO and the OLS regression equations and the OLS coefficient of determination R^2 are included; the 1:1 line (- - -) and the OLS line (- . - . -) are depicted

3.3 ET_o estimation in the absence of wind speed data

Two approaches were used, the first adopting the local annual average wind speed (u_{2avg}) as u_2 estimator, and the second using a default value of 2 m s^{-1} (u_{2def}) corresponding to the world average wind speed, adjusted for 3 m s^{-1}in case of weather stations located in airports because these show high average values because they are exposed to most of wind directions. In the Azores archipelago, there is a high spatial variability of the average wind speed (Table 5), even within the same island, which relates with the elevation and exposure to the dominant winds. For example, São Caetano and Santa Bárbara weather stations are located at high altitude (Table 1) but have small wind because they are leeward located in the mountains. Furnas and Granja stations are located in old volcanic craters and therefore less exposed to wind. Differently, Ribeirinha and Fontinhas stations are located on the top of small mountains exposed to winds and therefore show high u_2. For all study cases, u_2 is determined by the exposure to winds.

The accuracy of $ET_{o\ uavg}$ and $ET_{o\ udef}$ estimates, i.e., using respectively u_{2avg} and u_{2def} estimation procedures, is very high (Table 5, Fig. 10). $ET_{o\ uavg}$ shows no tendency for under- or over-estimation in most locations, with b_0 ranging 0.99 to 1.01 in 90% of cases and PBIAS ranging from −2.5 to 2.5% in most locations. R^2 values are very high ($R^2 > 0.94$) for all cases and RMSE values were small in all locations, below 0.28 mm day^{-1}. EF values were very high, with EF > 0.94 in all locations. Good results were also obtained for $ET_{o\ udef}$, where 95% of cases have b_0 ranging from 0.98 to 1.02 and all cases with $R^2 > 0.93$. RMSE was inferior to 0.36 mm day^{-1} at all locations and EF > 0.90 for all cases. Results of accuracy indicators show that results of $ET_{o\ uavg}$ and $ET_{o\ udef}$ are close to those of PM-ET_o, the variation of the latter is well explained by both ET_o approaches, the bias of estimation are small and vary locally, and high EF values were observed for both approaches indicating that the mean square errors were much smaller than the variance of PM-ET_o computed with full data sets. It is concluded that when u_2 data are not available, the most accurate procedure for computing ET_o consists of using u_{2avg} data, but these data are difficult to get for locations other than those used in this study. Nevertheless, the assessed accuracy of $ET_{o\ udef}$ was very good, close to the former; hence, the most useful approach consists in using u_{2def} in ET_o estimations.

Results in Fig. 11 show that dispersion of values around the regression line is higher for the windy locations, e.g., Ribeirinha, and, inversely, is small when the weather station is not exposed to the dominant high winds, e.g., S. Bárbara, with consequent impacts on R^2 values. For all cases, the regression coefficient is close to 1.0.

There are no results available for weather conditions similar to Azores. Various studies relative to humid or sub-humid climates using $u_2 = 2$ m s^{-1} report RMSE closer to ours (Sentelhas et al. 2010) or higher (Popova et al. 2006; Jabloun and Sahli 2008; Kwon and Choi 2011).

Table 5 Accuracy of ET_o estimations in the absence of wind speed data using the local annual average u_{2avg} and the default $u_{2def} = 2$ m s^{-1} (adjusted in case of weather stations located on grasslands in airports)

Weather stations	u_2 value (m s^{-1})	b_0	R^2	RMSE (mm day^{-1})	PBIAS (%)	EF
Santa Cruz das Flores*	$u_{2avg} = 3.2$	1.01	0.96	0.23	− 0.5	0.96
	$u_{2def} = 3.0$	1.00	0.96	0.23	0.8	0.96
Horta*	$u_{2avg} = 3.8$	1.00	0.97	0.23	− 0.5	0.97
	$u_{2def} = 3.0$	0.98	0.97	0.24	4.3	0.96
São Caetano	$u_{2avg} = 3.0$	0.99	0.98	0.10	0.4	0.98
	$u_{2def} = 2.0$	0.99	0.98	0.12	0.3	0.98
Velas*	$u_{2avg} = 3.8$	1.02	0.95	0.21	− 2.1	0.95
	$u_{2def} = 3.0$	0.98	0.96	0.20	2.4	0.95
Santa Cruz da Graciosa*	$u_{2avg} = 3.9$	1.01	0.95	0.26	− 0.9	0.95
	$u_{2def} = 3.0$	0.98	0.95	0.28	2.9	0.94
Angra Heroísmo	$u_{2avg} = 2.8$	1.00	0.98	0.17	− 0.4	0.98
	$u_{2def} = 2.0$	0.98	0.98	0.20	2.8	0.97
Granja	$u_{2avg} = 2.6$	1.02	0.98	0.16	− 2.6	0.98
	$u_{2def} = 2.0$	1.01	0.99	0.14	− 1.7	0.99
Lajes*	$u_{2avg} = 5.1$	1.00	0.96	0.24	− 0.4	0.96
	$u_{2def} = 3.0$	0.93	0.93	0.36	7.2	0.90
Ribeirinha	$u_{2avg} = 4.5$	1.01	0.98	0.17	− 1.4	0.98
	$u_{2def} = 2.0$	0.98	0.96	0.23	1.2	0.96
Santa Bárbara	$u_{2avg} = 2.3$	1.01	0.99	0.10	− 1.8	0.99
	$u_{2def} = 2.0$	1.01	0.99	0.09	− 2.0	0.99
Chã de Macela	$u_{2avg} = 1.4$	1.00	0.99	0.07	− 0.1	0.99
	$u_{2def} = 2.0$	1.01	0.99	0.09	− 2.1	0.99
Lagoa do Fogo	$u_{2avg} = 2.1$	0.99	0.99	0.10	0.5	0.99
	$u_{2def} = 2.0$	0.99	0.99	0.10	0.3	0.99
Furnas	$u_{2avg} = 2.4$	1.00	0.98	0.13	− 0.6	0.98
	$u_{2def} = 2.0$	0.99	0.98	0.14	0.6	0.98
Ponta Delgada*	$u_{2avg} = 3.8$	1.00	0.97	0.22	− 0.3	0.97
	$u_{2def} = 3.0$	0.98	0.97	0.24	2.3	0.96
Santana	$u_{2avg} = 1.7$	1.00	0.96	0.21	− 0.2	0.96
	$u_{2def2} = 2.0$	1.01	0.96	0.22	− 2.4	0.95
Sete Cidades	$u_{2avg} = 2.2$	1.00	0.98	0.11	0.4	0.98
	$u_{2def} = 2.0$	0.99	0.98	0.11	0.9	0.98
Tronqueira	$u_{2avg} = 1.4$	1.00	1.00	0.05	0.2	1.00
	$u_{2def} = 2.0$	1.00	0.99	0.08	0.1	0.99
Maia	$u_{2avg} = 2.3$	1.01	0.94	0.28	− 2.3	0.94
	$u_{2def} = 2.0$	1.00	0.95	0.27	− 0.7	0.95
Praia Formosa	$u_{2avg} = 3.3$	1.01	0.97	0.16	− 1.1	0.97
	$u_{2def} = 2.0$	0.96	0.97	0.20	4.2	0.96
Fontinhas	$u_{2avg} = 4.7$	0.99	0.97	0.15	− 0.6	0.97
	$u_{2def} = 2.0$	1.02	0.95	0.21	− 3.3	0.94

*Weather stations located at airports

4 Conclusions

The approaches proposed by Allen et al. (1998) tested in the present study for ET_o estimation when observed data is missing yielded good results. The performance of the tested approaches in Azores island environments are closely linked with the very humid climate and are influenced by longitude, altitude, and exposure to dominant winds.

In absence of RH observations, the actual vapor pressure e_a could be very accurately estimated, with negligible over- or under-estimation when deriving T_{dew} from $T_{mean} - a_d$, i.e., using a calibrated T_{dew} adjustment coefficient. That coefficient

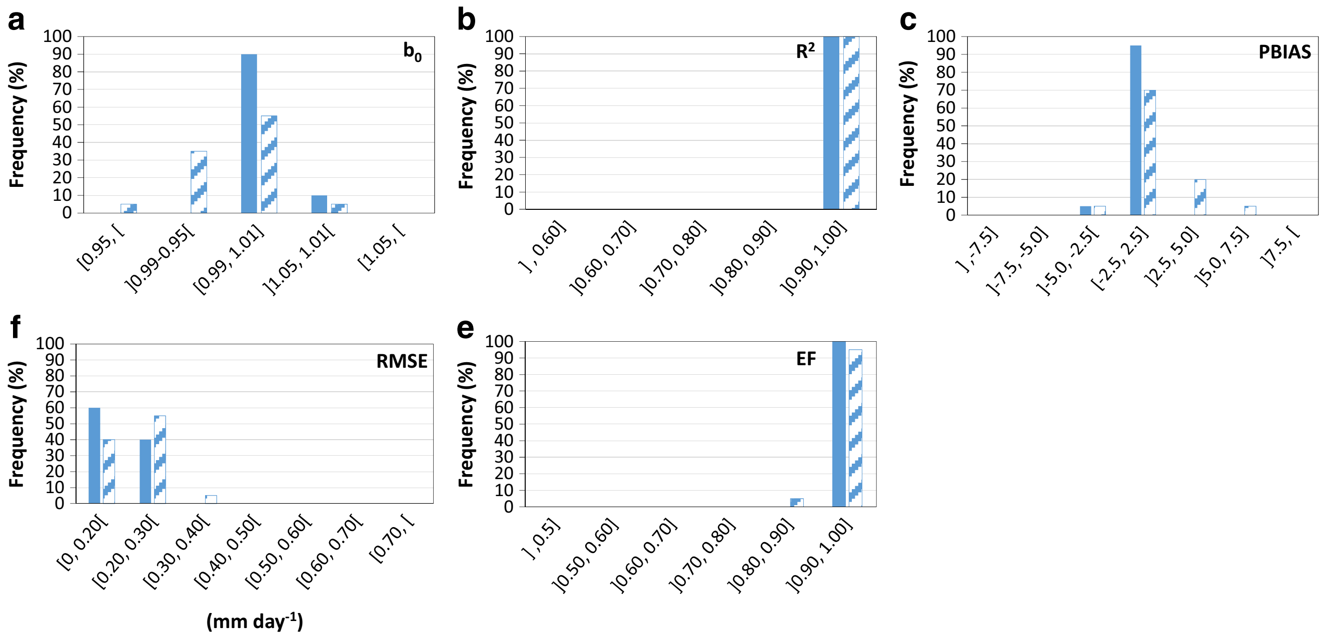

Fig. 10 Frequency distribution of the "goodness-of-fit" indicators relative to the computation of ET_o in the absence of wind speed using the local annual average, u_{2avg} ( ) and the default value $u_{2def} = 2$ m s^{-1} (or 3 m s^{-1} when adjusted for weather stations located on grasslands in airports) ( )

is larger in western islands, often is different in winter and summer, and is influenced by rainfall, elevation, and wind.

The use of a Hargreaves-based root squared temperature difference equation with a calibrated radiation adjustment coefficient k_{Rs} revealed a very good option to estimate both short-wave solar radiation, $R_{s\ TD}$, and ET_{oTD} when R_s observations are not available. k_{Rs} values were influenced by wind and by altitude of the locations, with lower values found in medium and high altitude locations. Despite tendencies for under-estimation of high R_s values and over-estimation of low ones, low estimation errors were obtained for R_s, with quite small over- or under-estimation errors of ET_o. Results reflect

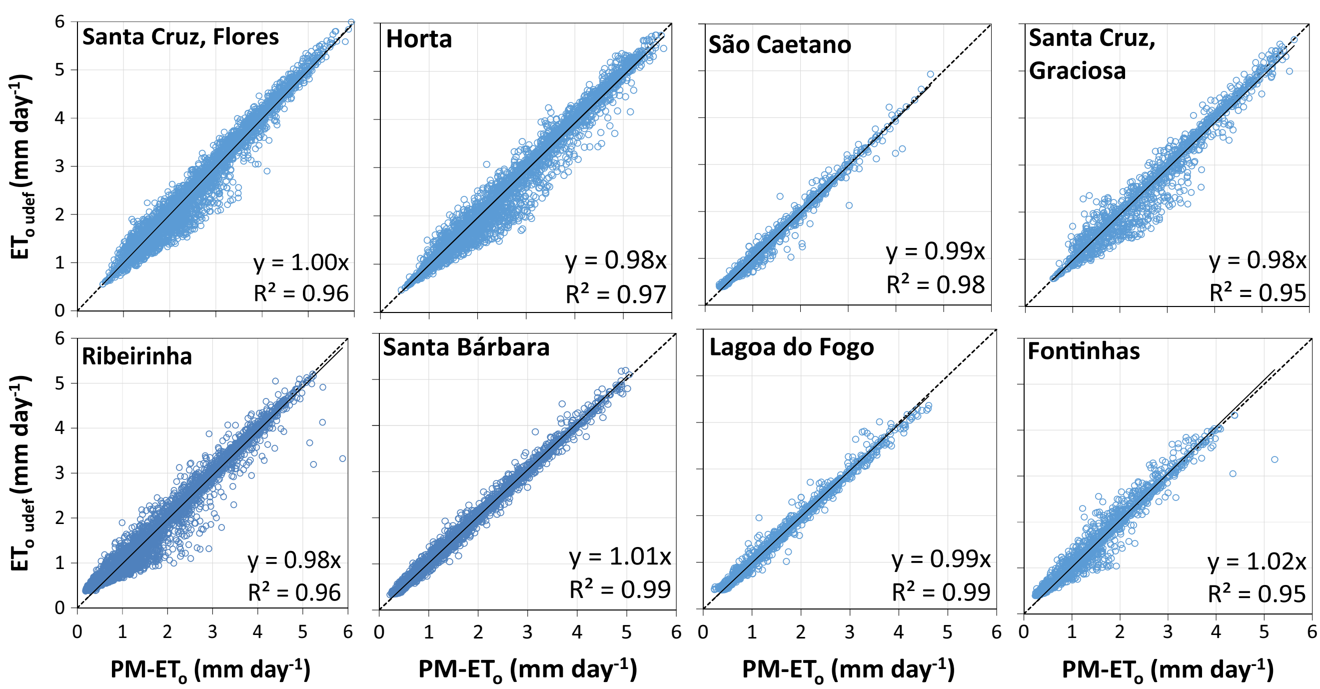

Fig. 11 Comparing $ET_{o\ udef}$ computed with the default wind speed with PM-ET_o using full data sets. The FTO regression equations and the OLS coefficient of determination R^2 are included; the 1:1 line is also depicted (- - -)

the high variability of R_s within the year and within the months due to frequent cloud cover, which highly reduces R_s in many rainy days. Nevertheless, errors of estimate were relatively small. It was found that k_{Rs} varies with altitude and wind, which influence the heat capacity of the air.

When observations of the wind speed are lacking, two options revealed appropriate to estimate u_2 to be used in the PM equation. The first is the use of the local annual average of u_2 and the second is the use of a default value, 2 m s^{-1}, which is majored for 3 m s^{-1} in case of weather stations located at airports because of the inherent exposure to all wind directions. Both cases, i.e., the use of the annual average of u_2 or a default value, yield very good accuracy indicators, with RMSE ranging 0.07–0.28 mm day^{-1}and EF > 0.90; however, it is recommended the use of the selected default values because they provide highly accurate estimates.

Results of this study provide for easy parameterization of the PM equation for reduced data sets in Azores archipelago. In particular, results of this study shall be used together with model CIELO to complement the respective spatial distribution of weather variables, which do not include R_s or ET_o while both may be computed using the variables yielded by the model.

The procedures tested herein are of great importance for humid and cold areas where temperature methods perform contradictorily (Trajkovic and Kolakovic 2009; Tabari 2010; Todorovic et al. 2013). Results obtained in the current study show that procedures herein assessed may be used for other cold humid locations, namely considering that the adjustment coefficients for both T_{dew} and $R_{s\ TD}$ estimations should be locally calibrated.

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