Testing the Rationale behind an Assumed Linear Relationship between Evapotranspiration and Land Surface Temperature

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Abstract: Theoretical considerations and empirical evidence indicate a linear relationship between the land surface temperature (T_s) and the corresponding evapotranspiration (ET) rate under spatially constant wind and net energy conditions at a homogeneous vegetated surface. Such a relationship lies at the core of the popular surface energy balance algorithm (SEBAL); the satellite-based energy balance approach for mapping evapotranspiration with internalized calibration (METRIC); and the lesser known calibration-free evapotranspiration mapping (CREMAP) method, just to name a few. The present findings are based on analytical solutions of the coupled turbulent heat and vapor transport equations and further corroborated by monthly reanalysis data of T_s, ET, and sensible heat transfer rates over extensive areas in North America and Europe, where the CREMAP method has previously been applied. **DOI: 10.1061/(ASCE)HE.1943-5584.0001091.**

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Introduction

Evapotranspiration (ET) is typically the dominant loss-term in terrestrial water balance calculations. With the continued rapid increase in global population and the corresponding dwindling of the available water resources on a per capita basis, any future catchment or regional-scale water resources planning depends heavily on the reliability of the ET estimates. Regional-scale planning involves accounting for the spatial variability of ET that traditional, small-scale measurements (i.e., eddy covariance and/or energy-balance Bowen ratio methods) cannot adequately provide for. With this in mind, remote sensing–based ET estimation approaches emerge naturally.

The literature of remote-sensing estimation of ET is vast because remote sensing data of land surface properties (e.g., surface temperature, vegetation status, albedo) became publicly available over the past 25 years, and the price of the relevant data storage quickly declined. For an exhaustive review of surface temperature–based ET estimation methods, see Kalma et al. (2008), or for a global view, see Wang and Dickinson (2012). There are many ways of classifying the existing ET estimation methods, see e.g., Kalma et al. (2008) for one possibility. Another possibility of classification could be based on the inherent key assumptions the methods use. For example, the studies of Su et al. (1999), Roerink et al. (2000), Jiang and Islam (2001), Venturini et al. (2004), Verstraeten et al. (2005), Wang et al. (2006), and Ma et al. (2014) exploit the self-preservation principle of the evaporative fraction, defined as the ratio of latent heat flux and available energy at the surface. Yet another possibility of classification could be based on the scale the ET estimation method is applicable (i.e., plot-scale, catchment/basin/regional-scale, or continental/global-scale), or whether it requires calibration of the parameters or not.

Recently Szilagyi et al. (2011a) and Szilagyi (2013b) proposed a simple regional-scale self-calibrating (thus being calibration-free) evapotranspiration (ET) mapping method (CREMAP) that transforms monthly daytime surface temperature (T_s) values obtained from moderate resolution imaging spectroradiometer (MODIS) measurements into corresponding ET rates. The approach is unique among the several existing basin-scale or regional-scale remote sensing–based ET estimation methods [for a review, see Senay et al. (2011)] in its self-calibrating nature and very modest input data requirement (i.e., T_s, air and dew point temperatures, incident global radiation or in lieu of it, sunshine duration or percent-possible sunshine); therefore, it merits further investigation of its inherent assumptions and limitations. Very recently Salvucci and Gentine (2013) also presented a (subdaily) calibration-free (i.e., self-calibrating) ET estimation method (based on different premises than CREMAP) that requires the same input as the present method with the friction velocity replacing the surface temperature. It can, however, be argued that it is easier to obtain remotely sensed daily or 8-day composited MODIS T_s values, as well as daily or monthly meteorological variables, than subdaily (30-min or hourly) values of the latter.

Short Review of the CREMAP Method

A central tenet of CREMAP is that the surface temperature values are directly proportional to the corresponding monthly ET values, because generally, evaporation cools the evaporating surface. It assumes that this proportionality between ET and T_s can be expressed by a temporally and regionally changing linear relationship. A necessary condition is that the net radiation (R_n) in a given month is near constant (within the region) over a flat or rolling vegetated terrain with quasi-homogeneous surface properties among (and not within) the MODIS cells of a spatial scale of approximately 1 km (or at a coarser spatial resolution). A prerequisite

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for a near-constant R_n term is a near-constant albedo value among the MODIS cells. It has indeed been the case in the two regions—Nebraska, in the United States and Hungary, in Europe—where CREMAP has already been applied and was validated with eddy covariance (Bugac, Matra, and Hegyhatsal of Fig. 1; Gothenburg, Odessa, and Mead of Fig. 2) and/or energy-balance Bowen ratio measurements (Sandhills of Fig. 2) on top of catchment water-balance data [Hidegviz, small experimental catchment, and Kapos, Marcal, Zala, and Zagyva, medium-sized watersheds of Fig. 1; USGS Hydrological Unit Code (HUC)-8 catchments of Nebraska and the Republican River basin (polygons), shared by Colorado, Kansas, and Nebraska, in Fig. 2]. In Mead, measurements were simultaneously performed over irrigated and rainfed corn [Oak Ridge National Laboratory Distributed Active Archive Center (ORNL DAAC) 2013]. In the Sandhills, the measurement sites included a subirrigated meadow, a dry meadow, and an upland location (Billesbach and Arkebauer 2012).

In temperate regions in general, with good vegetation cover and/or relatively high soil moisture content, MODIS albedo values typically express a low spatial variance between the MODIS cells with a standard deviation of approximately 10% of their nominal monthly values for the March–November period (Kovacs 2011; Szilagyi et al. 2011a). A 10% albedo value change means only a few percent change in the net shortwave radiation balance when the surface has a typical albedo value of approximately 0.15 (i.e., $0.85R_d$ versus 0.865 or $0.835R_d$, where R_d is the downward shortwave radiation at the surface). Patchy vegetation cover combined with low soil moisture near the surface can deteriorate the linear relationship between surface temperature and ET because the vegetation may cool itself effectively through access to soil moisture with the help of a deep root-system; but this temperature is confused by the surrounding warmer soil temperature (e.g., Moran et al. 1997). In fact, under such conditions and over subdaily intervals, higher surface temperatures may result in higher ET rates. Monthly averaging, however, evens out such possible irregularities to a certain extent.

In the winter months, when patchy snow cover is likely (and thus, the spatially near-constant R_n requirement is grossly violated), no such transformations were recommended in the CREMAP approach. Then, cell ET rates are assumed equal to the regional ET rate provided by the WREVAP model of Morton et al. (1985) based on the complementary relationship (CR) (Bouchet 1963) of evaporation. The latter takes into account the complicated feedback mechanisms between the evaporating surface and the overlying air. The CR relates the difference in ET rates of plot-size versus regional-scale wet surfaces (meaning a spatial extent equal or larger than approximately 1 km) (Brutsaert 1982; Morton et al 1985), to actual evapotranspiration. Plot-size wet-surface ET rates can be larger than that of a regional scale owing to local energy advection from the surrounding drier and hotter land with reduced ET. The larger this reduction, the stronger the horizontal energy transfer, and so the difference between the two wet-surface ET rates (Bouchet 1963).

In a recent study, McMahon et al. (2013) concluded that CR-based ET estimation methods are the most reliable among existing practical ET estimation approaches. For a detailed description of CREMAP, see Szilagyi et al. (2011a) or Szilagyi (2013b). Figs. 3 and 4 display validation of CREMAP with the help of the energy-balance Bowen ratio (Sandhills sites), eddy covariance, and water balance data. In Gothenburg, Odessa (Landon et al. 2009) and Hegyhatsal, the EC energy balance is not closed; therefore, the applied corrections are in the form of constant multipliers, e.g., Szilagyi et al. (2011a), for more detail. The flux stations have a footprint in the order of a few hundred meters, except at Hegyhatsal, where the footprint is in the order of ~10 km because of the 130-m elevation

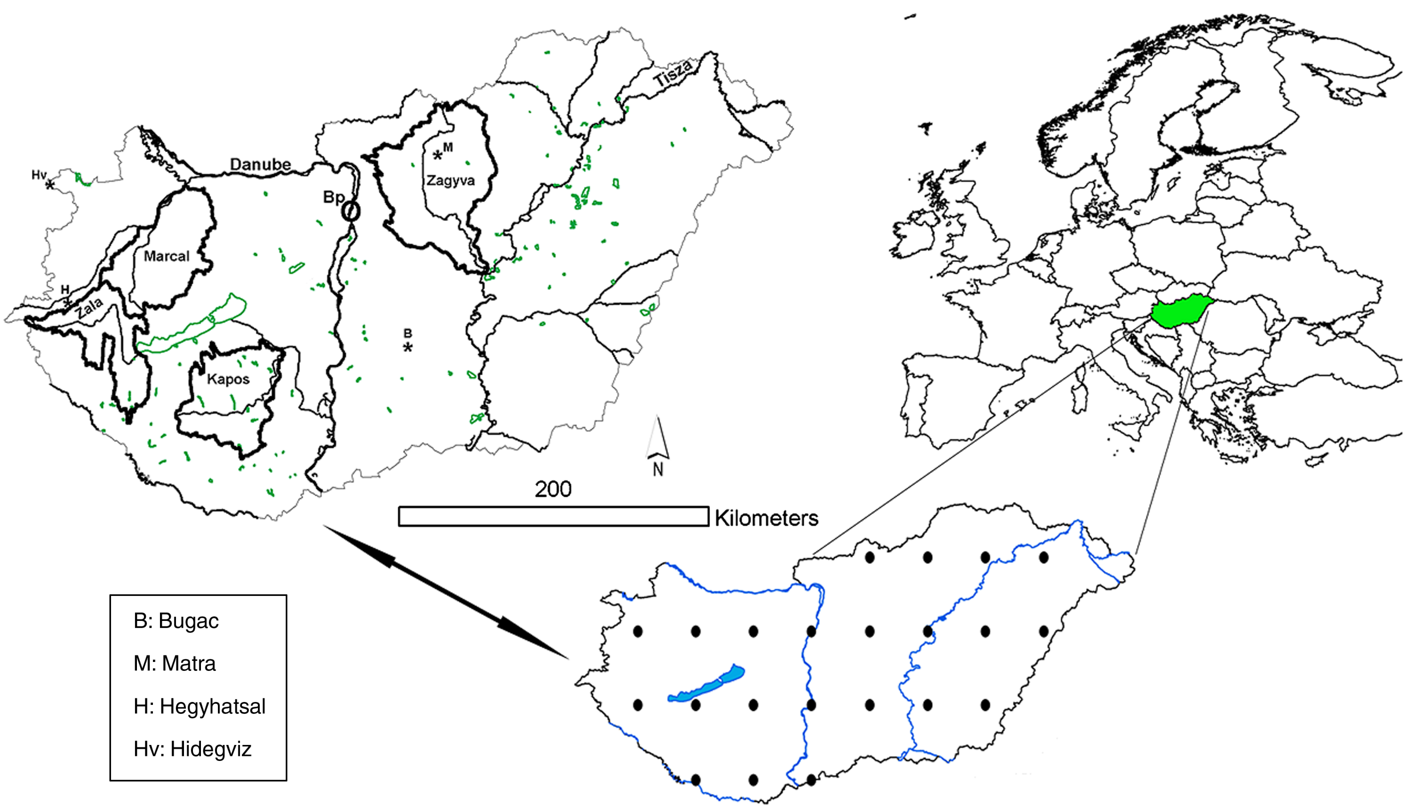

Fig. 1. Locations where the CREMAP method has been validated in Hungary; the dots denote the ERA-Interim cell centers of a spatial resolution of 0.7°

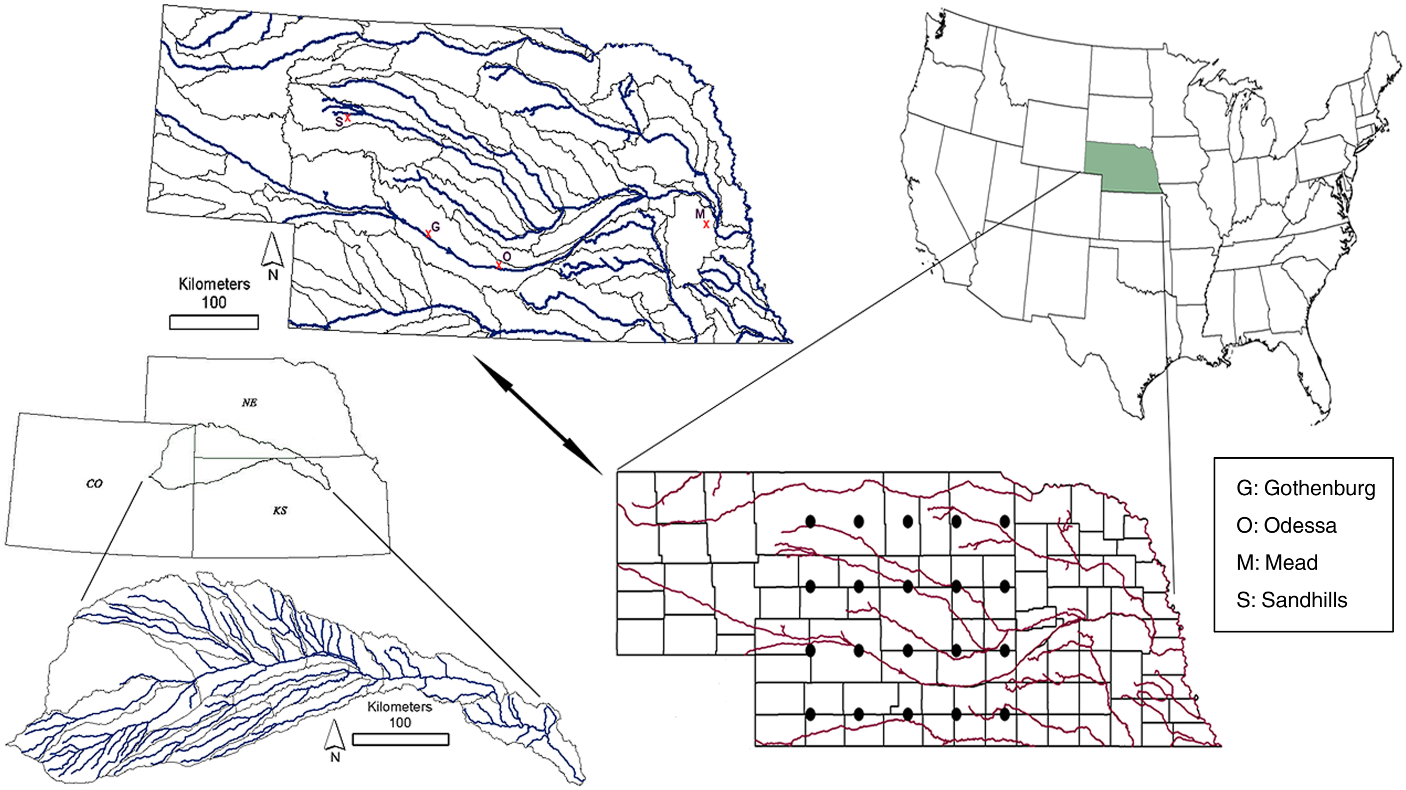

Fig. 2. Locations where the CREMAP method has been validated in Nebraska; the dots denote the ERA-Interim cell centers of a spatial resolution of 0.7°

of the instruments above the ground. In Fig. 4, P is precipitation, Ro is stream discharge, ΔGW is the change in groundwater elevation (primarily attributable to large-scale irrigation) expressed in water-depth equivalent, and S_y is the specific yield. Specific yield values were given as intervals (therefore, the high, mean, and low distinctions) for different aquifer materials (Cederstrand and Becker 1998). See Szilagyi and Jozsa (2013) and Szilagyi (2014) for more detail.

Additional practical water resources applications of the method are found in Szilagyi (2013a, 2014), Szilagyi and Jozsa (2013), and Szilagyi et al. (2011b, 2012, 2013).

Objectives

Although a linear relationship between the T_s and ET values is explicitly defined in the CREMAP approach, such a relationship only implicitly appears in the popular plot-scale ET estimation methods of SEBAL (Bastiaanssen et al. 1998) and METRIC (Allen et al. 2007) when net radiation at the surface is constant. Recently Bateni and Entekhabi (2012) also demonstrated a strong dependence of energy partitioning into sensible and latent heat fluxes on the land surface temperature.

The objective of this study is the investigation of (1) whether such an assumed linear relationship between ET and T_s can be derived theoretically; and also (2) if it can be validated with published, independent, regional-scale modeled ET values. The latter are provided by the European Centre for Medium-Range Weather Forecasts (ECMWF) in the form of monthly ET estimates based on reanalysis data. The theoretical derivation is based on a simplified version of the coupled 2D turbulent vapor and heat transport equations (Yeh and Brutsaert 1971). Reanalysis data combine measurements with model simulations and are considered to be the best representation of reality. In this study, the existence of a linear relationship between the surface temperature and the corresponding ET rate represented by modeled (in lieu of measured) ET values that are not based on such an assumption is to be established. Furthermore, the necessary conditions when such a relationship can indeed be expected are defined. These may strengthen confidence in ET estimation methods that use (implicitly or explicitly) such an approach.

Surface Temperature versus ET Relationship from the Coupled 2D Turbulent Heat and Vapor Transport Equations

Following Yeh and Brutsaert (1971) and Brutsaert (1982), one may consider a sudden moisture and temperature change at the otherwise homogeneous land surface perpendicular to the prevailing mean horizontal wind, $\bar{u}$, blowing along the x-direction of a Cartesian coordinate system (Fig. 5). Because everything is assumed to be homogeneous along the y-direction of the horizontal plane, the y and z components of the mean wind vector can be considered zero without loss of generality. By denoting with K_v and K_h the vertical tensor components of the turbulent diffusivity for vapor and heat; and assuming that they are comparable, one can write $K_v \approx K_h = K$. This equality is called Reynold's analogy [see Brutsaert (1982) for historical background and further explanation] and is often evoked in turbulence studies. From a first-order closure approach (i.e., the so-called K-theory) for the turbulent fluxes, the coupled 2D steady vapor and heat transport equations become (Yeh and Brutsaert 1971)

$$\bar{u}\frac{\partial \bar{q}}{\partial x} = \frac{\partial}{\partial z}\left(K\frac{\partial \bar{q}}{\partial z}\right) \qquad \bar{u}\frac{\partial \bar{T}}{\partial x} = \frac{\partial}{\partial z}\left(K\frac{\partial \bar{T}}{\partial z}\right) \tag{1}$$

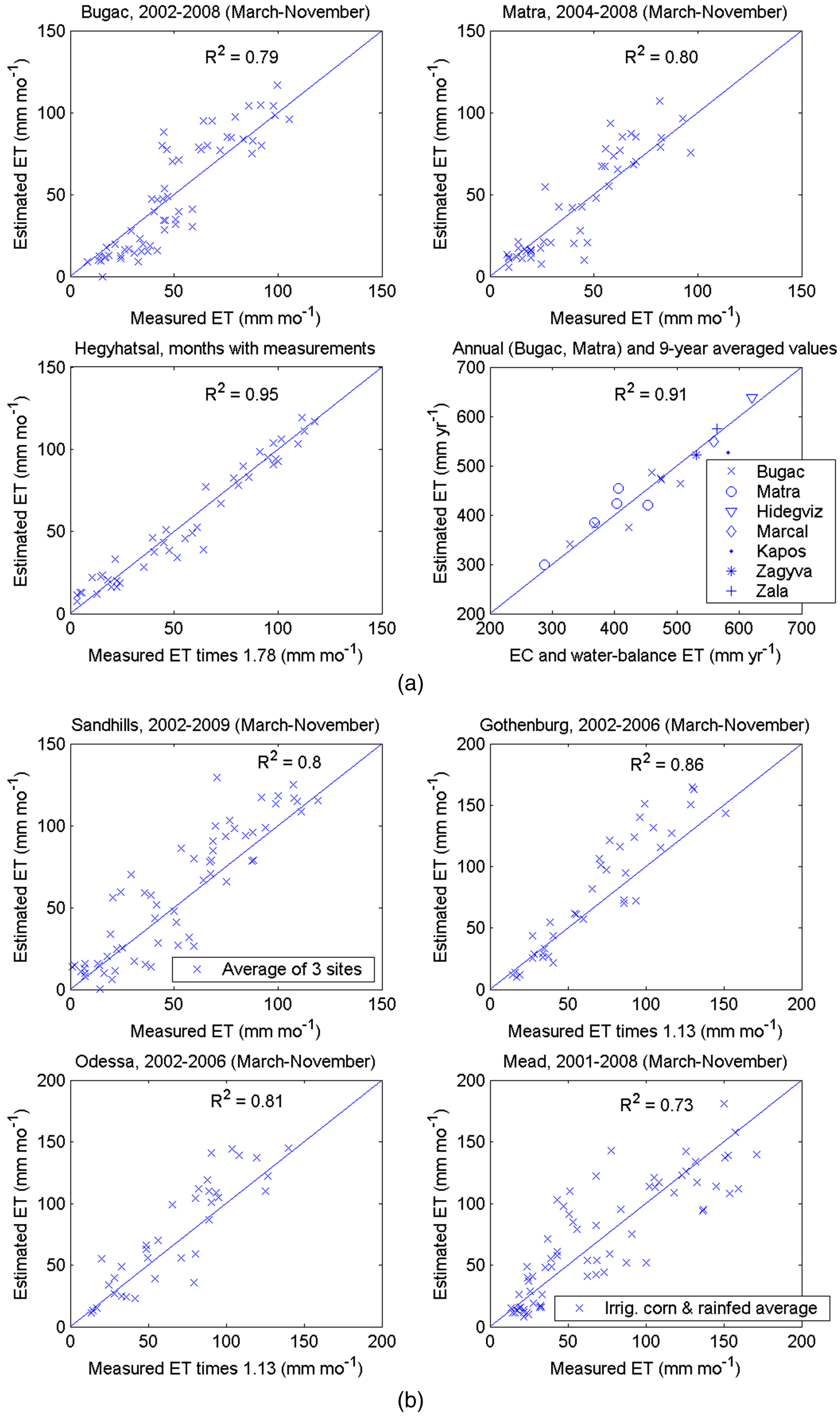

Fig. 3. Results of the CREMAP model validation in (adapted from Szilagyi et al. 2011a): (a) Hungary (for locations see Fig. 1); (b) Nebraska (flux stations only; see Fig. 2); R^2 is the explained variance

where $\bar{q}$ = mean specific humidity; $\bar{T}$ = mean potential temperature to be substituted by the mean air temperature owing to the close proximity to the land surface. The prescribed equilibrium $\bar{u}(z) = az^m$ and $K(z) = bz^n$ profiles are assumed to remain unchanged across the sudden moisture change. The recent power-function formulation of the two profiles is preferred to the classical logarithmic approach owing to better differentiation properties near the surface (Brutsaert and Yeh 1970). From experimental data, one obtains $a = 5.5u_* z_0^{-m}(7m)^{-1}$; $b = u_* z_0^m(5.5m)^{-1}$; and $n = 1 - m$ (Yeh and Brutsaert 1971), where u_* is the friction velocity, and z_0 is the roughness height of the surface, but they are not needed to be defined this specific way for the solution of Eq. (1).

By allowing the specific humidity, $\bar{q}_a(z)$, and temperature, $\bar{T}_a(z)$, profiles upwind of the surface-property change to remain

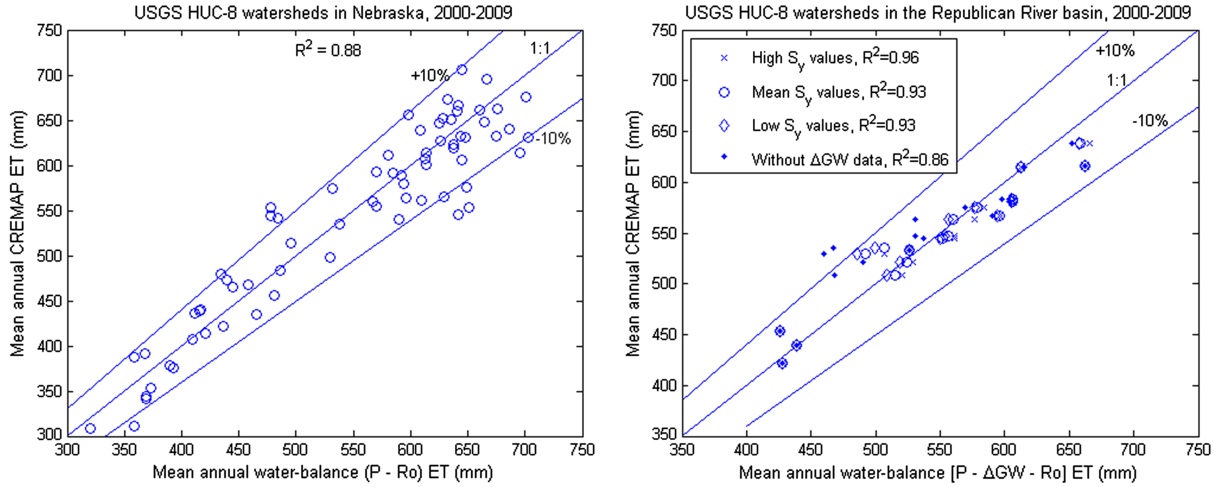

Fig. 4. Water balance validation results of the CREMAP model for USGS HUC-8 watersheds in Nebraska and in the Republican River basin; R^2 is the explained variance (adapted from Szilagyi 2013a; Szilagyi 2014)

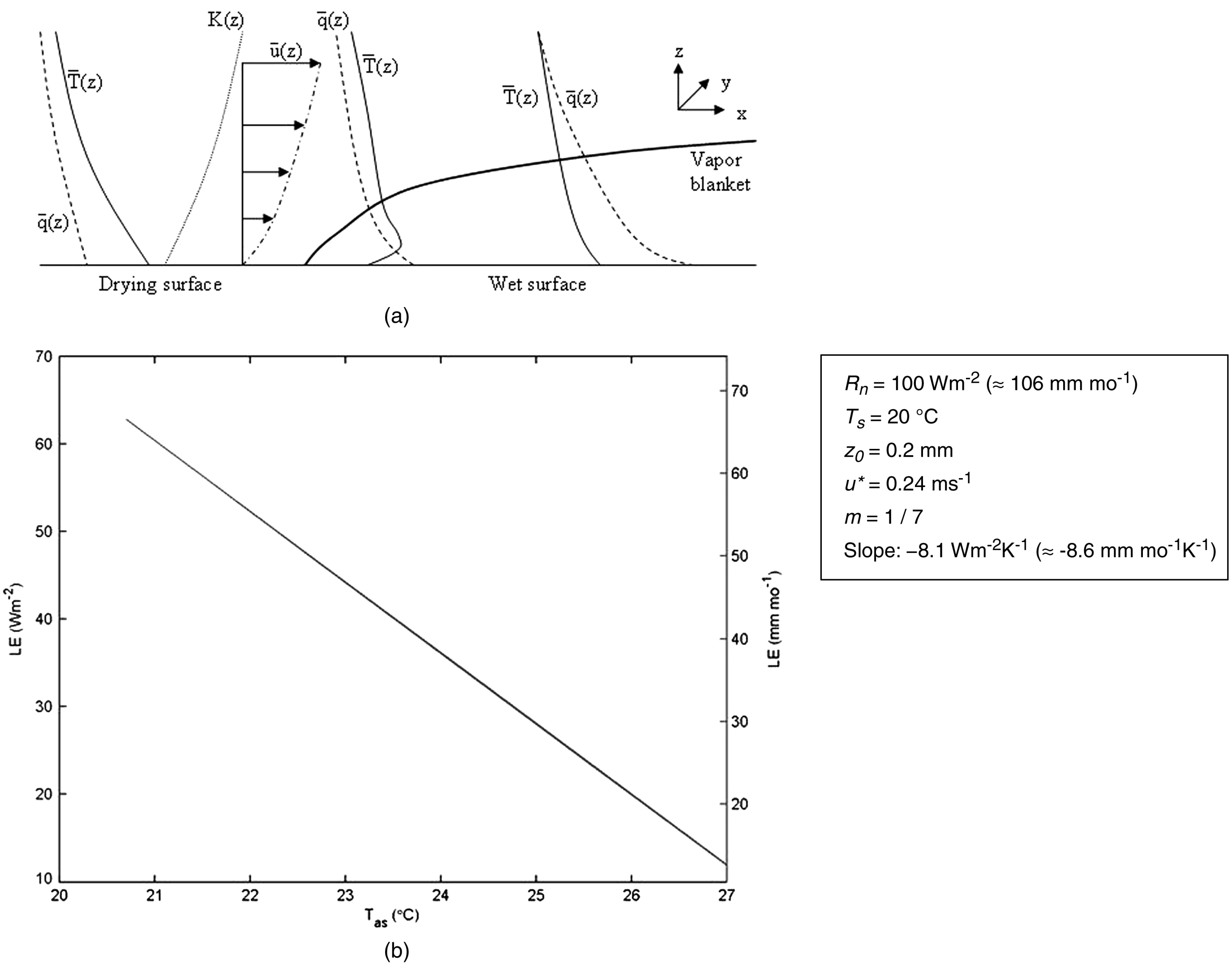

Fig. 5. (a) Schematics of the heat and vapor transfer described by Eq. (1); (b) The analytical solution–derived linear relationship between the drying land surface temperature, T_{as}, and the corresponding latent heat transfer, LE

in an equilibrium (the subscript a refers to the nonwet arid conditions), as the initially uniformly wet surface dries out, one can write

$$\frac{\partial}{\partial z}\left(K\frac{\partial \bar{q}_a}{\partial z}\right) = \frac{\partial}{\partial z}\left(K\frac{\partial \bar{T}_a}{\partial z}\right) = 0 \tag{2}$$

Eq. (2) expresses that the latent and sensible heat fluxes (proportional to the terms within the parentheses) are constant along the vertical direction near the surface. Integration of Eq. (2) in the vertical direction between the land surface and elevation, z, yields the two profiles as

$$\bar{T}(z) = T_s - \frac{H}{c_p b\rho(1-n)} z^{1-n} \qquad \bar{q}(z) = q_s - \frac{E}{b\rho(1-n)} z^{1-n} \tag{3}$$

written now for the initially uniform wet surface of temperature T_s and corresponding (saturated) specific humidity q_s. The sensible heat (H) and vapor (E) fluxes found in Eq. (3) are defined as $H = -c_p \rho K \partial\bar{T}/\partial z$ and $E = -\rho K \partial\bar{q}/\partial z$, respectively. When Eq. (3) is applied for the drying equilibrium profiles, then T_s is replaced by T_{as} and q_s by (unsaturated) q_{as}. In this case, ρ = density of air and c_p = specific heat of air at constant pressure. The equilibrium assumption is justified for mean conditions over a day or longer periods. By considering heat conduction to the ground to be negligible over such periods, the sum of the sensible and latent heat fluxes equals the temporally constant net radiation term, R_n, at the surface. Yeh and Brutsaert (1971) demonstrated that the solution of Eq. (1) for $H(x)$ and $E(x)$ along the wet patch under a constant R_n term [and unchanged $\bar{u}(z)$) and $K(z)$ profiles] yields the so-called Sutton (1934) solution (not specified here for sake of simplicity) (e.g., Yeh and Brutsaert 1971; Brutsaert 1982, for a more detailed description) with the following equations for the surface temperature and specific humidity values:

$$T_s = T_{as} - \frac{L_e(q^*_{as} - q_{as})}{c_p + \alpha_q L_e} \qquad q_s = q_{as} + \frac{c_p(q^*_{as} - q_{as})}{c_p + \alpha_q L_e} \tag{4}$$

where the star denotes the saturated value of the specific humidity at the drying surface temperature; L_e = latent heat of vaporization; and α_q = slope of the saturated specific humidity curve at an intermediate temperature between T_s and T_{as}. T_s does not depend on the x-coordinate value, unlike the H and E terms, which implies that the wet patch maintains a constant uniform surface temperature (and thus a constant q_s) under a constant R_n term (Szilagyi and Jozsa 2009a, b). From Eq. (4)

$$\frac{T_s - T_{as}}{q_s - q_{as}} = -\frac{L_e}{c_p} \tag{5}$$

can be obtained, relating the drying land surface temperature to its specific humidity value (Szilagyi and Jozsa 2009a, b).

A linear relationship between T_{as} and the heat fluxes, $LE(= L_e E)$ or H, immediately follows because Eq. (5) is linear for q_{as} once T_{as} is specified (T_s and therefore q_s are supposed to be known), as is Eq. (3) with q_{as} and T_{as} supplied for the drying land surface and evaluated at an elevation above the ground, where the temperature and humidity changes are negligible during the drying of the land (Dyer and Crawford 1965; Rao et al. 1974). The (unchanging) $\bar{T}$ and $\bar{q}$ values at that elevation can be obtained from Eq. (3) applied for the initial, uniformly wet condition once H and LE are specified from the $H + LE = R_n$ and $H/LE = \gamma/\Delta(T_s)$ equations, in which, γ is the psychrometric constant and Δ is the slope of the saturation vapor pressure curve evaluated at T_s.

Fig. 5 displays the resulting linear relationship with $R_n = 100\ \mathrm{Wm^{-2}}$ (in water equivalent equals approximately 106 mm mo^{-1} for the average month of 30.25 days); $T_s = 20°\mathrm{C}$; $z_0 = 0.2$ mm; $u_* = 0.24\ \mathrm{ms^{-1}}$; $m = 1/7$ (the latter three values from Yeh and Brutsaert 1971); and an elevation $z = 100$ m at which the drying land surface has a supposedly negligible impact. The slope of the line is approximately $-8.1\ \mathrm{Wm^{-2}K^{-1}}$ ($\approx -8.6\ \mathrm{mm\,mo^{-1}K^{-1}}$).

The preceding derivation of obtaining a linear relationship between T_{as} and LE involved certain assumptions [i.e., homogeneous surface, constant R_n, unchanging $u(z)$ and $K(z)$ profiles] that are likely to be violated in the more complex physical environment. Therefore, it is necessary to validate the analytical result, demonstrated in Fig. 5, with independently obtained model LE values that do not build on a surmised linear relationship.

Surface Temperature versus ET Relationship from ERA-Interim Reanalysis Data

The ERA-Interim reanalysis data of ECMWF is produced by an improved atmospheric model and data assimilation system, replacing those previous models that produced ERA-40 data. ERA-Interim data in this study have a spatial resolution of 0.7 degree. It covers both study areas, Nebraska and Hungary (Figs. 1 and 2), from which monthly mean reanalysis values of the surface temperature, as well as monthly sums of the sensible (H) and latent heat (LE) transfer terms were extracted for the 2000–2010 period. For a detailed description of how H and LE were obtained, see the ECMWF (2007) documentation.

Both study areas are represented by a relatively mild topography, with a predominantly agricultural landscape interspersed with forested regions. The continental climate of both regions is characterized by a mean annual temperature of ~10°C and annual precipitation of approximately 600 mm.

In the present study, it was assumed that soil heat conduction is negligible for a vegetated land surface in the summer months (June, July, August); therefore, the sum of H and LE was assumed equal to R_n in each month studied. The large cell size (0.7 degree) of the ERA-Interim grid (which the T_s, H, and LE values are representative of) combined with a monthly time step assure that surface properties among the cells (including the surface albedo) are quasi-constant in a given month over the study areas (note the absence of large mountains). For Nebraska, an area with limited extent was chosen (Fig. 2) because the land surface gradually rises more than 1 km from east-to-west across the state, resulting in potentially large differences in atmospheric, radiation, and surface conditions.

Fig. 6 displays the ERA-Interim monthly ET and R_n rates for Nebraska in June within the 2000–2010 period. Widespread and very intensive large-scale irrigation in July and August may affect the ET rates at a regional scale (Szilagyi 2013b), unfavorably reducing their range for plotting the ET versus T_s value pairs. In June, irrigation is still negligible within the state of Nebraska, resulting in an inverse ERA-Interim ET versus T_s linear relationship: as ET decreases, T_s increases [Fig. 6(a)]. In Fig. 6(a) the sharp bracket denotes the population mean, and r is the linear correlation coefficient.

By defining a narrow interval for R_n, i.e., $150 \pm 1.5\ \mathrm{mm\,mo^{-1}}$ (when expressed in water equivalent) to pick cells with R_n values falling into this range [Fig. 6(b)], it could be checked if a linear relationship indeed exists between ET and the corresponding surface temperature of the ERA-Interim cells when $R_n \approx$ constant. The resulting relationship is significant ($R^2 = 0.64$), corresponding to a linear correlation coefficient value (r) of -0.8 [Fig. 6(c)] with a slope of approximately $-4.6\ \mathrm{mm\,mo^{-1}K^{-1}}$. Circles come from months when R_n was within 2%, whereas crosses are within

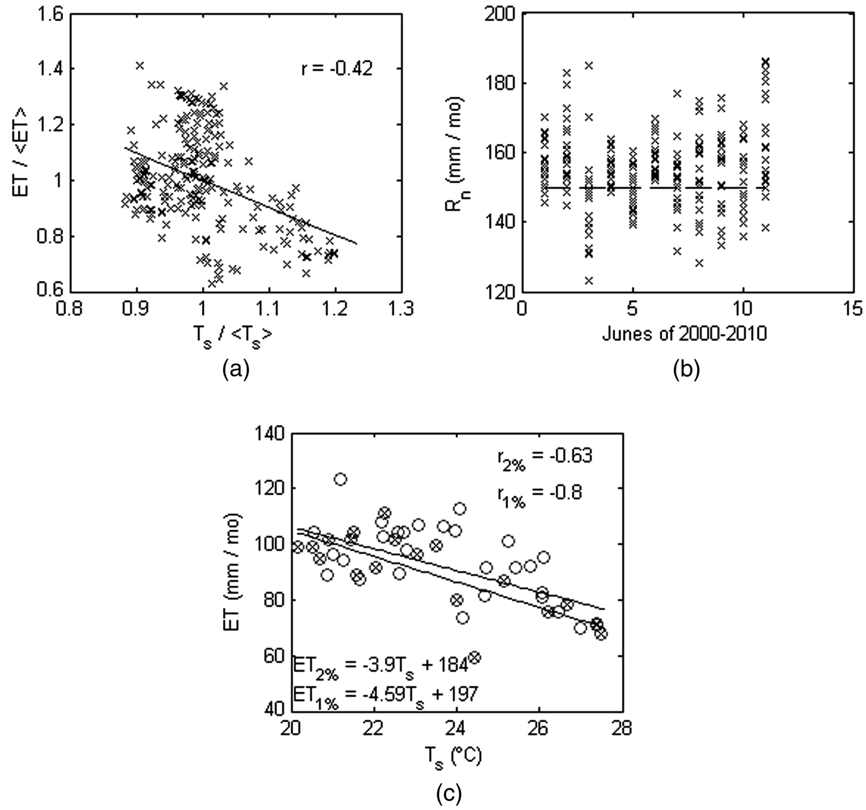

Fig. 6. ERA-Interim monthly net radiation (R_n) and ET rates within Nebraska for June 2000–2010: (a) normalized cell ET versus T_s values scattered around the best-fit first-order polynomial line; (b) net radiation at the surface (horizontal line denotes $R_n = 150\ \mathrm{mm\,mo^{-1}}$); (c) ET rate versus T_s in months having $R_n = 150\ \mathrm{mm\,mo^{-1}}$

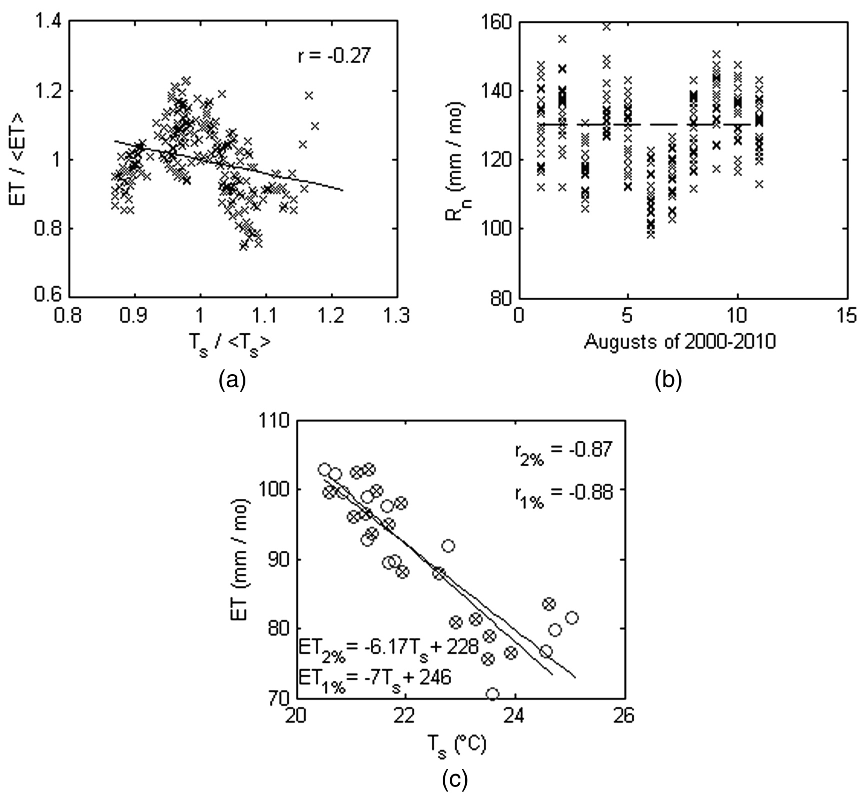

Fig. 7. ERA-Interim monthly net radiation (R_n) and ET rates within Hungary for August 2000–2010: (a) normalized cell ET versus T_s values scattered around the best-fit first-order polynomial line; (b) net radiation at the surface (horizontal line denotes $R_n = 130\ \mathrm{mm\,mo^{-1}}$); (c) ET rate versus T_s in months having $R_n \approx 130\ \mathrm{mm\,mo^{-1}}$

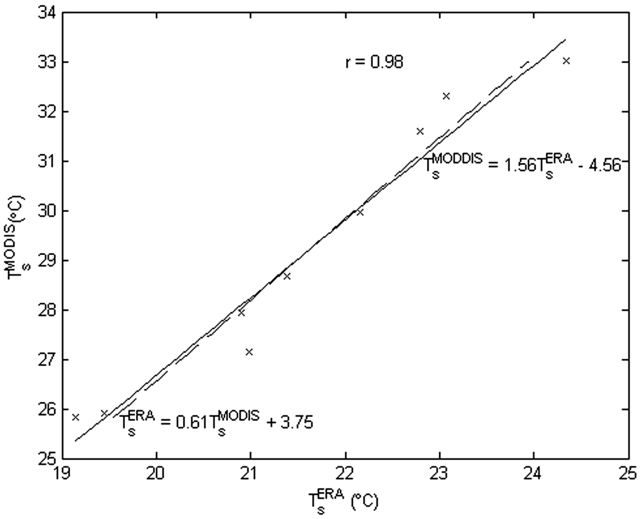

Fig. 8. Linear correlation between the ERA-Interim and MODIS T_s values (averaged over Hungary), month of August, 2000–2008, the period of available MODIS values (Kovacs 2011); lines are the best-fit first-order polynomials; r denotes the linear correlation coefficient value

1% of 150 mm mo^{-1}. The lines are the best-fit first-order polynomials, with the corresponding correlation coefficients, $r_{2\%}$ and $r_{1\%}$, respectively.

The ET versus T_s relationship strengthens for Hungary ($R^2 = 0.77$), having an r value of -0.88 for R_n in the range 130 ± 1.3 mm mo^{-1} (Fig. 7) for the month of August, with a corresponding slope of -7 mm mo^{-1}K^{-1}. August is the driest summer month in Hungary, expressing the largest range in the T_s values (largely depending on how much rain took place in the wetter months of June and July) and yielding the strongest correlation ($R^2 = 0.96$, $r = 0.98$) between the ERA-Interim and MODIS derived surface temperature values (Fig. 8). In Hungary, crop-irrigation is negligible. Notably, the MODIS T_s values used in CREMAP are daytime surface temperatures, whereas the ERA-Interim values are representative of the entire 24-h day. The strong linear relationship between the two types of surface temperature values (at least at a monthly time step and regional scale) supports the transferability of the results obtained with one type of T_s to the other.

The ERA-Interim data at a monthly level support the theoretically derived linear relationship between the surface temperature and the corresponding ET rate under near-constant surface net radiation. The resulting linear relationships are stronger for Hungary than for Nebraska. A likely explanation is that regional advection of energy is more substantial in the windy and widely open High Plains of Nebraska than in the much calmer Carpathian basin, containing Hungary and surrounded by substantial mountains from every direction. The slope of the $ET_{1\%}$ versus T_s curve for Hungary (i.e., -7 mm mo^{-1}K^{-1}) is also close to the theoretical value of -8.6 mm mo^{-1}K^{-1} in Fig. 5. Moreover, the average slope of the CREMAP-applied ET versus T_s^{MODIS} transformations for August is -5.28 mm mo^{-1}K^{-1} (Kovacs 2011), which becomes -8.38 mm mo^{-1}K^{-1} with the application of ERA-Interim surface temperature values from the T_s^{ERA} versus T_s^{MODIS} equation of Fig. 8.

Having an ERA-Interim data derived slope of -7 mm mo^{-1}K^{-1} for the T_s versus ET transformation close to the CREMAP-applied -8.38 mm mo^{-1}K^{-1} value suggests that the T_s versus ET relationship is not only linear under a constant net surface energy term, but also that its month-to-month construction within CREMAP is likely adequate. The construction involves choosing two anchor points each month: the regional mean T_s value with the corresponding regional ET rate specified by Morton's WREVAP program (Morton et al. 1985), and the wet-environment ET rate, obtained from the Priestley-Taylor equation (Priestley and Taylor 1972), with the matching coldest pixel surface temperature value.

Summary

This study is the first that, through theoretical considerations and experimental data, validated the frequently applied, but never fully justified, linear relationship that is used, e.g., in such models as SEBAL (Bastiaanssen et al. 1998), METRIC (Allen et al. 2007), and SSEB (Senay et al. 2007), beside CREMAP, between the surface temperature and the corresponding ET rate, when net energy at the surface is constant. Although in all of the aforementioned models both the coldest and warmest pixels must be chosen for the identification of the linear transformations, CREMAP requires a regional mean of the surface temperature in place of the warmest pixel. This seems like a minor but nonetheless important step in obtaining more accurate and stable linear transformations, because in temperate regions, the warmest pixels may be far from being completely dry (i.e., to have a presumed negligible LE rate), an important assumption in those (i.e., SEBAL, METRIC, or SSEB) models. Also, the warmest pixels may have (i.e., when the standard deviation of the T_s values is large) an associated albedo value significantly different from the rest of the pixels, bringing into question the resulting linear relationship between T_s and ET that requires, as it was demonstrated in this study, a quasi-constant net energy term.

Because CREMAP is a calibration-free approach with minimal data requirement, it is recommended for regional-scale ET estimation purposes in combination with MODIS daytime surface temperature data in areas where measured ET fluxes are not available for calibration. The method is best suited to be applied at a monthly time step over vegetated land surfaces of a flat or rolling terrain. The finest temporal resolution the CREMAP method can be applied with confidence is approximately 5 days due to its reliance on the WREVAP-calculated regional ET rate. WREVAP is based on the complementary relationship of Bouchet (1963) that builds upon an equilibrium state between the land and the ambient atmosphere. The time needed to reach such equilibrium may take several days (Morton et al. 1985). Similarly, the finest spatial resolution the CREMAP method is expected to work reliably is approximately 1 km, the size of the MODIS cells. It is so because the smaller the spatial scale, the larger the expected inhomogeneity of the land surface, i.e., the more the cells will differ in, e.g., albedo and surface roughness values. Thus, the surface homogeneity assumption of the method becomes seriously violated.

Besides the previously mentioned studies of CREMAP-aided recharge estimation, the ET estimation method has recently been applied in runoff modeling (Szilagyi 2013b) and in regional-scale water balance calculations (Szilagyi 2014). Although in general, surface temperature-based ET estimation methods work with a relative error of 15–30% (Kalma et al. 2008), CREMAP estimates are typically within 10–15% of water-balance or flux measurement derived ET rates (Szilagyi 2013b, 2014; Szilagyi et al. 2011a).

Acknowledgments

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