Intercomparison of Evapotranspiration Estimates at the Different Ecological Zones in Jordan

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ABSTRACT

An estimate of evapotranspiration (ET) is needed for many applications in diverse disciplines such as agriculture, hydrology, and meteorology. The objective of this study was to compare two methods for estimating daily actual ET ($ET_a$) from six sites located in four different ecological zones within Jordan. The first method used the analytical land–atmosphere radiometer model (ALARM) and the dimensionless temperature procedure, whereas the second method used $ET_a$ calculated from the FAO-56 reference evapotranspiration. ALARM converts general remotely sensed surface temperatures to aerodynamic temperature. Standard meteorological data from weather stations were used with both methods, and the Moderate Resolution Imaging Spectroradiometer (MODIS)–based leaf area index, surface temperature, and albedo were obtained to estimate $ET_a$, using the former method. A validation study was conducted on an alfalfa field in Jordan Valley using ALARM and the American Society of Civil Engineers’ (ASCE) method, which is very similar to FAO-56 except it uses alfalfa as a reference crop. Because this alfalfa field was irrigated and because of warm air advection, ET rates based on measurements of soil moisture change ranged from about 6 to 10 mm day$^{-1}$. For this range, the root-mean-square error (RMSE) for ALARM was 0.87 mm day$^{-1}$ and the coefficient of determination $r^2$ was 0.36, whereas the RMSE for ASCE was 1.25 mm day$^{-1}$ and $r^2 = 0.06$. There was good agreement between minimum, maximum, and average $ET_a$ for the two methods at all sites except for Irbid, for which the minimum and, consequently, the average were different. Much of the site-to-site and temporal variability was found to be statistically significant. Reasons for this variability include soil types, vegetation cover, irrigation, and warm advection.

1. Introduction

Estimates of evapotranspiration (ET) are needed for many applications in diverse disciplines, such as agriculture and hydrology. Many studies of long-term averages have shown that more than half of the net solar energy, and subsequently two thirds of precipitation, goes to ET (Brutsaert 1982). ET is linked to the land surface energy budget as follows (e.g., Brutsaert 1982):

$$R_n - G = H + E,$$

where $R_n$ (W m$^{-2}$) is the net incoming radiation, $G$ is the heat flux into the ground (W m$^{-2}$), and $H$ (W m$^{-2}$) and $E$ (W m$^{-2}$) are the sensible and latent (evaporative) heat fluxes, respectively, into the atmosphere. For the energy balance to close, any part of $R_n - G$ that does not contribute to $E$ must be converted into $H$. For that to happen, the surface has to have a temperature ($T_s$) that forces the energy balance to close. The estimation of $H$ (or ET as a residual) over vegetated terrain is based on an aerodynamic temperature ($T_i$), which is the temperature that gives the correct value of $H$ at a specified value (denoted $z_{oh,j}$) of the scalar roughness length $z_{oh}$ based on the Monin–Obukhov similarity (MOS) theory in the surface sublayer (Brutsaert 1982; Stull 1988). Specification of the value of $z_{oh}$ to give the correct value of $H$ for use with a radiometric surface temperature ($T_r$) is a difficult problem (e.g., Mahrt and Vickers 2004); Crago and Suleiman (2005) outlined a method (discussed here in section 2a) to
specify $z_{0h,i}$ and to convert $T_r$ to $T_i$. In the MOS theory, the flux is proportional to the difference between $T_i$ and air temperature ($T_a$), with the ratio $H/(T_i - T_a)$ depending on variables characterizing the atmospheric turbulence and the land surface. This relationship can be expressed as (e.g., Brutsaert 1982)

$$H = \frac{(T_s - T_a)ku_a\rho}{\ln\left(\frac{z_a - d_0}{z_{0h}}\right) - \psi\left(\frac{z_a - d_0}{L}\right)}$$

(2)

where $T_s$ (°C) is the surface temperature, $T_a$ (°C) is the air temperature at a height $z_a$ (m) in the surface sublayer, $k$ (where $k = 0.4$) is von Kármán’s constant, $u_a$ (m s$^{-1}$) is the friction velocity, $\rho$ (kg m$^{-3}$) is the density of the air, $c_p$ (J kg$^{-1}$ K$^{-1}$) is the specific heat at constant pressure, $z_{0h}$ (m) is the scalar roughness length for sensible heat, and $d_0$ (m) is the displacement height. Atmospheric stability, which affects the efficiency of turbulent transport, is included by means of $\psi$, which is a function of the stability or buoyancy parameter $(z_a - d_0)/L$, where $L$ (m) is the Monin–Obukhov length. Once $T_s$ is known, it can be applied to calculate $H$ [Eq. (2)] and then actual ET can be obtained as a residual [Eq. (1)]. In one example, the accuracy of regional scale actual ET [obtained as a residual from (1) after finding $H$ using (2), with $T_s$ measured at a single point within the region] is approximately 70%–80% (Wang et al. 2005). Models have been developed to improve accuracy through the use of radiometric surface temperature (Hatfield et al. 1983; Ben-Asher et al. 1992; Kustas et al. 2007; Anderson et al. 2007), leaf area index (LAI; Consoli et al. 2006), and net radiation (Bandara 2003) available from visible and infrared bands of satellite data. One example of ET modeling from remote sensing data is the Surface Energy Balance Algorithm for Land (SEBAL), which uses an empirical relationship between radiometric surface temperature and the difference between aerodynamic surface temperature and air temperature for each pixel (Bastiaansen et al. 1998). Several studies (e.g., Menten and Choudhury 1993; Bastiaansen and Chandrapala 2003; Chandrapala and Wimalasuriya 2003; Allen et al. 2005; Tasumi et al. 2005; Wang et al. 2005) used and modified SEBAL for spatial estimates of ET with remote sensing and weather data.

To implement SEBAL to estimate ET, it must be possible to identify a wet pixel and a dry pixel within an area of interest, and it must be reasonable to assume that atmospheric conditions aloft are horizontally uniform over that area. When one or both of these conditions cannot be met, the surface energy balance index (SEBI) can be used to calculate relative ET using a ratio of temperature differences (Roerink et al. 2000). The minimum surface temperature difference is obtained by solving the similarity equation for the minimum sensible heat flux that is found as a residual after determining the potential ET. The maximum surface temperature difference is determined by assuming that ET is zero. Because these bounds are pixel dependent, potential ET has to be calculated for each pixel.

Suleiman and Crago (2004) developed a dimensionless temperature, $\Delta_T$ that does not require information from wet and dry pixels nor potential ET. For each pixel, the maximum surface temperature is determined by assuming that ET for the pixel is zero, and the minimum temperature is determined by assuming that sensible heat flux is zero. This approach is advantageous in practice when a dry pixel is not available (Qiu et al. 2006). The variable $\Delta_T$ can be defined as $(T_i - T_a)/(T_{\text{max}} - T_a)$, where $T_i$ is the aerodynamic surface temperature, $T_a$ is the air temperature, and $T_{\text{max}}$ is the surface temperature that would occur if all the available energy $(R_n - G)$ were converted to $H$ and no evaporation occurred. The dimensionless temperature procedure was used for mapping ET at a local scale, with hydrological applications at riparian meadow restoration sites in California (Loheide and Gorelick 2005).

The analytical land–atmosphere radiometer model (ALARM) has been developed to convert the radiometric surface temperature ($T_r$) to the aerodynamic surface temperature ($T_i$) at any view angle (Crago 1998; Suleiman and Crago 2002a). ALARM converts radiometric surface temperature, measured at any view angle, to a well-defined aerodynamic surface temperature ($T_i$) by correcting for vegetation temperature profile and considering LAI, canopy height, fractional cover, leaf angle distribution, and sensor zenith view angle. ALARM worked well for varied canopy density when the zenith view angle was less than 20° and worked satisfactorily for view angles greater than 20° (Suleiman and Crago 2002b; Zibognon et al. 2002). Other models such as Lhomme et al. (2000) and Massman (1999) also worked best at near-nadir view angles (Suleiman and Crago 2002a).

The objective of this study was to compare the estimated daily actual ET from different ecological zones in Jordan using two methods. The first method used ALARM and the dimensionless procedure, whereas the second method used the actual ET calculated from FAO-56 reference ET ($E_{TR}$). The first method relied on LAI, albedo, and $T_r$ obtained from the Moderate Resolution Imaging Spectroradiometer (MODIS) Terra. In addition, both methods used climatic parameters obtained from weather stations. Although MODIS data have been used for evapotranspiration...
estimation (Allen et al. 2007; Senay et al. 2007), the study might be considered as the first step in the application of MODIS data and the dimensionless temperature approach in estimating actual ET in Jordan and similar environments. Also, this work is an advance in technique because it is the first application of the ALARM method coupled to the evapotranspiration estimate given in Suleiman and Crago (2004). The most significant contribution of this work is ALARM because it converts general remotely sensed surface temperatures to aerodynamic temperature. This is an important topic because methods for estimating actual evapotranspiration from satellite data have remained elusive principally because of the problem of how to make use of remotely sensed surface temperatures.

2. Theoretical background

a. ALARM description

Within ALARM, the foliage is assumed to have an exponential vertical temperature profile (Brutsaert and Sugita 1996) as follows:

\[ T_f = T_{f_0} + (T_{f_h} - T_{f_0})e^{-h \xi}, \]

where \( T_f \) is the temperature of the foliage at a height \( z \) above the soil surface; \( T_{f_0} \) is the foliage temperature at the top of the canopy; \( T_{f_h} \) is the asymptotic limit of the exponential foliage temperature profile, far below the bottom of the canopy; \( \xi = (h - z)/h \) is the dimensionless depth into the canopy; \( h \) is the canopy height; and \( b \) is a decay constant. Qualls and Yates (2001) observed an exponential vertical temperature profile within a grass canopy.

ALARM converts radiometric surface temperature measured at any view angle to a well-defined aerodynamic surface temperature \( T_i \) by correcting for vegetation temperature profile and considering LAI, canopy height, fractional cover, leaf angle distribution, and zenith sensor view angle as follows:

\[ T_i = T_i + (T_{f_h} - T_{f_0})(w - W), \]

where \( W \) is defined below in (7), and \( w \) can be derived (Crago 1998) as

\[ w = (1 - f_{s_0}e^{-b})(\frac{\mu_s b}{g'LAI} + 1)^{-1}. \]

In (5), \( f_{s_0} = \exp[-g'(\text{LAI})/\mu_s] \) is the fraction of soil seen by the IRT (Friedl and Davis 1994); \( \mu_s \) is the cosine of the view zenith angle; and \( g' \) is taken as 0.5, which corresponds to a spherical leaf angle distribution and is representative of a wide range of vegetation types. When \( T_{f_h} = T_{f_0} \), the canopy is isothermal. Under these conditions, Brutsaert and Sugita (1996) showed that the resulting scalar roughness length \( z_{0h,i} \) is given by

\[ z_{0h,i} = z_0 \exp \left[ \frac{h}{(h - d_0)r_2} + \ln \left( \frac{h - d_0}{z_0} \right) \right], \]

where \( d_0 \) is zero plane displacement height, \( z_0 \) is momentum roughness length, and \( r_2 \) is defined in (2) to estimate the correct \( H \) using the \( z_{0h,i} \) for \( z_{0h} \). Alternatively, \( T_i \) is the temperature the surface would require to give the correct sensible heat flux if the canopy were isothermal. The \( r_2 \) is given by \( r_2 = \frac{[a - (a^2 + 4C_2)^{1/2}]/2}{(a - (a^2 + 4C_2)^{1/2})/2} \) and in (4), \( W \) is

\[ W = -(r_2 + b)C_2[\frac{r_2(b^2 + ba - C_2)}{2}]. \]

In turn, \( C_2 = 2\text{(LAI)}(C_t)h/[k(h - d_0)] \), and \( C_t \) is the transfer coefficient in the bulk transfer equation for the foliage elements, given by \( C_t = C_r Re_s^{mPr^{-n}} \). The variable \( a \) is an exponential decay parameter of eddy diffusivity, \( Pr \) is the Prandtl number, and the Reynolds number that is appropriate for transport through a leaf boundary layer is \( Re_s = u_{\theta}L_f/\nu \), where \( L_f \) is the characteristic length scale of a leaf and \( \nu \) is the kinematic viscosity.

In (5), \( w \) is a weighting coefficient, describing the importance of \( T_{f_h} \) and \( T_{f_0} \) in determining the radiometric surface temperature seen by a radiometer,

\[ T_r = wT_{f_h} + (1 - w)T_{f_0}. \]

Similarly, \( W \) is a weighting coefficient, describing the relative importance of \( T_{f_h} \) and \( T_{f_0} \) in producing sensible heat flux.

b. ALARM parameterization

ALARM has several variables \( (T_{f_h}, b, a, \text{and } d_0/h) \) that need to be parameterized, all of which have real physical meanings independent of the means of measuring surface temperature. Crago and Suleiman (2005), in a study of different sites with varying LAI, found that the use of a generalized parameterization for these variables at the different sites gave sensible heat flux values comparable to those obtained using localized parameterization. Based on their findings, they recommended the following generalized parameterization for the four variables \( (T_{f_h}, b, a, \text{and } d_0/h) \):

\[ T_{f_h} = T_a, \]

where \( T_a \) is the air temperature at the top of the canopy.
The parameter $b$ controls the rate at which foliage temperature increases with depth into the canopy and was parameterized as a function of LAI:

\[ b = 0.75 \text{ for LAI} \geq 1.87, \quad \text{and} \quad b = 3.7 - 1.58\text{LAI} \text{ for LAI} < 1.87. \] (10)

Previous work with ALARM (Zibognon et al. 2002; Suleiman and Crago 2002a,b) suggests that the parameters $a$ and $d_0/h$ can influence the estimates of ET. Specifically, larger values of $a$ (near 5) effectively confine turbulence and turbulent transport to the top layers of the model canopy, whereas smaller values (near 0) allow turbulence. They were parameterized as follows:

\[ a = 0.5\text{LAI} \quad \text{and} \quad d_0/h = 0.335a. \] (12)

\[ d_0/h = 0.335a. \] (13)

c. Dimensionless temperature

Suleiman and Crago (2004) introduced a dimensionless temperature ($\Delta T$) as follows:

\[ \Delta T = \left( \frac{T_s - T_u}{T_{\text{max}} - T_u} \right). \] (14)

A value of $T_{\text{max}}$ can be obtained by solving for $T_s$ in Eq. (2), assuming that $H$ equals $R_n - G$. The relationship between $H$ and $\Delta T$ is approximately linear and goes through the origin (assuming the denominator of (2) varies little as $T_s$ goes from $T_i$ to $T_{\text{max}}$):

\[ H = (R_n - G)\Delta T. \] (15)

The relationship between $E$ and $\Delta T$ is

\[ E = (R_n - G)(1 - \Delta T), \] (16)

and the evaporative fraction EF is

\[ \text{EF} = \frac{E}{(R_n - G)} = 1 - \Delta T = \frac{T_{\text{max}} - T_s}{T_{\text{max}} - T_u}. \] (17)

The dimensionless temperature $\Delta T$ is obtained from (14) using ALARM $T_u$, measured $T_r$, and $T_{\text{max}}$ can be found as described above. Scaling $T_r$ using the dimensionless procedure, reduces the sensitivity of ET estimates to errors in $T_u$ and $T_r$. The assumption of a constant evaporative fraction, $\text{EF} = E/R_n$, was implemented to extend instantaneous to daily ET because orbiting satellites usually provide coverage only once a day.

In all, there are three major assumptions in the integrated ALARM and dimensionless algorithms. Within ALARM, the foliage is assumed to have an exponential vertical temperature profile. This assumption should be generally valid during the satellite pass in the middle of the day because exponential foliage vertical temperature profiles are more evident in the middle of the day. The relationship between $H$ and $\Delta T$ is assumed to be linear and goes through the origin. This assumption should not result in any serious errors in the estimation of ET, especially because the variation of the denominator of (2) is little as $T_s$ goes from $T_i$ to $T_{\text{max}}$. The assumption of a constant evaporative fraction contributes to uncertainty in ET estimates, but these uncertainties should be minimal most of the time in semiarid climatic conditions.

d. FAO-56 evapotranspiration

Daily actual ET was obtained from the FAO-56 reference ET ($\text{ET}_0$) approach to compare it with the ALARM actual ET. Allen et al. (1998, 2005) emphasized that the FAO-56 Penman–Monteith (PM) reference ET ($\text{ET}_0$) would provide the best estimate of ET under various climatic conditions. The FAO-56 $\text{ET}_0$ was developed for a hypothetical well-watered and actively growing uniform grass height of 0.12 m, with a surface resistance of 70 s m$^{-1}$ and an albedo of 0.23 (Allen et al. 1998). The equation for a grass reference crop according to Allen et al. (1998) is defined as follows:

\[
\text{ET}_0(\text{mm day}^{-1}) = \frac{0.408(R_n - G) + \frac{900}{T + 273}u_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34u_2)},
\] (18)

where $R_n$ is the net radiation (MJ m$^2$ day$^{-1}$), $G$ the soil heat flux (MJ m$^2$ day$^{-1}$), $T$ the mean daily air temp (°C), $u_2$ the mean daily wind speed at 2 m height (m s$^{-1}$), $e_s - e_a$ the saturation vapor pressure deficit (kPa), $\Delta$ the slope of the vapor pressure–temperature curve (kPa °C$^{-1}$), and $\gamma$ the psychometric constant (kPa °C$^{-1}$). The terms in the numerator on the right side of the equation are available energy and air dryness forcing, respectively (Kim and Entekhabi 1997). Actual ET, $\text{ET}_u$, is found from $\text{ET}_0$ as $\text{ET}_u = K_cK_s\text{ET}_0$, where $K_c$ is the crop coefficient, and $K_s$ is a water stress coefficient (Allen et al. 1998).
3. Site description and data

Two separate studies were conducted, first to validate the models used and then to apply the models to test sites to understand the dynamics of evaporation in an arid-to-semiarid region. The test sites will be described first and will discuss the geography and climate of the region.

a. Study area and selected sites

Jordan is located between 29°11’ and 33°22’N latitude and between 34°19’ and 39°18’E longitude, with an area of more than 89 000 km². More than 80% of the country’s area is arid and receives less than 100 mm annual rainfall, with the precipitation pattern being latitude, longitude, and altitude dependent. Rainfall decreases from north to south, from west to east, and from higher altitudes to lower altitudes. Average rainfall ranges from 600 mm yr⁻¹ in the north to less than 50 mm yr⁻¹ in the south and east. The rainy season is between October and May, with 80% of the annual rainfall occurring between December and March. During the rainy season, most of precipitation is orographic, resulting from the passage of frontal depressions across the Mediterranean near Cyprus. Frequent droughts occur in many parts of the country and usually result in the failure of rainfed agriculture. Therefore, the mapping of actual ET and vegetation response to rainfall would provide an invaluable tool to assess spatial and temporal distribution of droughts and to formulate real-time action plans (Ministry of Environment 2006).

According to Al-Eisawi (1982), the country may be divided into four ecological zones or bioclimatic regions (Fig. 1; Al-Bakri and Suleiman 2004).

- Mediterranean: restricted to the highlands of Jordan, with altitudes ranging from 700 to 1750 m above mean sea level and mean annual rainfall ranging from 300 to 600 mm. This region supports the best natural
vegetation in Jordan including forest stands. In addition to natural vegetation, rainfed cultivation of wheat and other field crops, summer crops, and orchards is practiced.

- **Irano–Turanian (also known as steppe):** surrounds the Mediterranean region from all sides except the north and is characterized by no forest cover. Altitude ranges from 500 to 700 m and annual rainfall ranges from 100 to 300 mm. Vegetation is mainly dominated by low shrubs and bushes (timberless land) and some rainfed barley cultivation.

- **Saharo–Arabian:** comprises most of the country, known as the Badia, with altitudes of 600–700 m and a mean annual rainfall of less than 100 mm. Dry, hot summers and relatively cold, dry winters characterize this region. The region is classified as rangeland and is home to a wide range of highly adapted organisms.

- **Sudanian Penetration:** provides unique ecosystems, with altitudes varying from the lowest elevation on earth at 400 m below the sea level (at the Dead Sea in the Rift Valley) up to 1200 m in the south. This region is characterized by very hot summers and warm winters, with a mean annual rainfall of 50 mm or less. Vegetation is dominated by *Acacia sp.* in the low-altitude region and scattered shrubs in the high-altitude region.

Six weather stations distributed within different ecological zones in Jordan (Fig. 1) were used in this study. The selection of stations was restricted by the availability of data. Although more than 30 stations are operating in the country, daily climatic records are only available for a few stations. In 2002, daily climatic data were only available for the six weather stations that were included in this study. The Rwaished site was located within the Saharo–Arabian region. The Safawi and Mafraq sites were located in the Irano–Turanian region. The Aqaba site was located within the Sudanian Penetration and the Irbid and Amman sites were located within the Mediterranean region. Monthly precipitation for the study sites is shown in Table 1. The high variation of monthly and total rainfall amounts from one site to another is apparent in Table 1.

Land use/cover of each site was derived from the visual interpretation of high-resolution Landsat Enhanced Thematic Mapper Plus (ETM+) images, using an onscreen digitizing procedure for each of the 1 km × 1 km pixels containing the study sites. Output from land use/cover mapping of each site was used to calculate the weighted average of water stress and crop coefficients, $K_s K_c$, shown in Table 2, which was required to convert ET$_0$ to actual ET.

### Data

Eight-day 1-km LAI, albedo, and instantaneous 4-km radiometric surface temperature ($T_r$) were obtained from the MODIS Terra instrument for April 2002 for the six sites. To compensate for any possible positional shift, a first-order geometric correction was made for the MODIS images of albedo and LAI using a geo-coded image of Advanced Very High Resolution Radiometer (AVHRR). Output images were resampled using the bilinear interpolation, and the digital numbers of the relevant data were extracted for the test sites. Albedo, the adjusted reflectance value at the mean solar zenith angle for a 16-day period, was derived from the spectral reflectance data (MOD43B) of the first seven bands using the conversion coefficients described by Strahler et al. (1999).

Linear interpolation was used to obtain daily LAI from the eight-day LAI. In an earlier study of bare and sparsely vegetated areas, Wan et al. (2002) found that the MODIS land surface temperatures (LSTs) agreed well with in situ measured LSTs within ±1°C in the range −10°C to 49°C for both daytime and nighttime LST; similar uncertainties likely apply here as well. The use of MODIS data for ET predictions does not require site or species-specific calibration (Nagler et al. 2005). Although a functional relationship between LAI and plant canopy height ($h$) is vegetation dependent, $h$ was estimated from LAI as $h (m) = 0.2$ LAI when LAI was

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<th></th>
</tr>
</thead>
<tbody>
<tr>
<td>Rwaished</td>
<td>32.1</td>
<td>16.2</td>
<td>28</td>
<td>5.9</td>
<td>3.2</td>
<td>0.3</td>
<td>85.7</td>
</tr>
<tr>
<td>Safawi</td>
<td>28.9</td>
<td>17.6</td>
<td>37.7</td>
<td>12.8</td>
<td>5.7</td>
<td>0.3</td>
<td>103</td>
</tr>
<tr>
<td>Mafraq</td>
<td>19</td>
<td>65.2</td>
<td>81.4</td>
<td>13.9</td>
<td>39</td>
<td>17.2</td>
<td>235.7</td>
</tr>
<tr>
<td>Irbid</td>
<td>51.2</td>
<td>205.1</td>
<td>133.3</td>
<td>40.4</td>
<td>117.7</td>
<td>65.8</td>
<td>617.5</td>
</tr>
<tr>
<td>Amman</td>
<td>20.3</td>
<td>93.8</td>
<td>109.8</td>
<td>29.3</td>
<td>38.4</td>
<td>21.2</td>
<td>314.2</td>
</tr>
<tr>
<td>Aqaba</td>
<td>9.3</td>
<td>1</td>
<td>3</td>
<td>4.4</td>
<td>0</td>
<td>0</td>
<td>17.7</td>
</tr>
</tbody>
</table>

greater than or equal to 0.5, and $h$ (m) = 0.1 when LAI was less than 0.5. The range of LAI and $T_r$ at the Mafraq site includes 70.3% of protected rangeland ($K_sK_c = 0.20$), whereas Aqaba includes bare sandy soils with a very sparse vegetation ($K_sK_c = 0.05$).

Table 2. Land use/cover percent, water stress $\times$ crop coefficient ($K_sK_c$), and the weighted $K_sK_c$ for each site.

<table>
<thead>
<tr>
<th>Continuous urban</th>
<th>Green urban</th>
<th>Rainfed crops</th>
<th>Irrigated olives</th>
<th>Rangeland*</th>
<th>Bare rock</th>
<th>Weighted $K_sK_c$</th>
</tr>
</thead>
<tbody>
<tr>
<td>Rwaished</td>
<td>38.0</td>
<td>9.5</td>
<td>—</td>
<td>5.5</td>
<td>47.0</td>
<td>0.22</td>
</tr>
<tr>
<td>Safawi</td>
<td>29.6</td>
<td>7.4</td>
<td>—</td>
<td>1.7</td>
<td>51.1</td>
<td>0.17</td>
</tr>
<tr>
<td>Mafraq</td>
<td>3.6</td>
<td>0.9</td>
<td>—</td>
<td>—</td>
<td>95.5</td>
<td>0.19</td>
</tr>
<tr>
<td>Irbid</td>
<td>20.2</td>
<td>5.0</td>
<td>67.8</td>
<td>7.0</td>
<td>—</td>
<td>0.80</td>
</tr>
<tr>
<td>Amman</td>
<td>57.5</td>
<td>6.4</td>
<td>—</td>
<td>21.2</td>
<td>14.9</td>
<td>0.30</td>
</tr>
<tr>
<td>Aqaba</td>
<td>32.9</td>
<td>16.1</td>
<td>—</td>
<td>—</td>
<td>51.0</td>
<td>0.19</td>
</tr>
<tr>
<td>$K_sK_c$</td>
<td>0.00</td>
<td>1.00</td>
<td>1.00</td>
<td>1.00</td>
<td>0.15</td>
<td>0.00</td>
</tr>
</tbody>
</table>

* Mafraq site includes 70.3% of protected rangeland ($K_sK_c = 0.20$), whereas Aqaba includes bare sandy soils with a very sparse vegetation ($K_sK_c = 0.05$).

The 2001/02 season was wet, with few cold days (minimum air temperature of 4.8°C at Mafraq) and some hot days (maximum air temperature of 34.5°C at Aqaba and Rwaished) throughout the end of April. According to the official climatic records (Jordan Meteorological Department 2002; available online at http://www.jmd.gov.jo), the 2001/02 rainfall season was above average for all sites except for Aqaba. Irbid had the highest precipitation amounts in April, then Amman Airport, and then Mafraq, whereas the other three stations received 0 or about 0 rainfall (Table 1). Wind speed, $u$, was measured at a height of 2 m using Dines’s pressure tube anemographs 4 times a day, 0000, 0600, 1200, and 1800 local time. The $u$ at 1200 was used in this study because it was the closest to the time of Terra MODIS passes (1000–1200) over the six sites.

Hourly and daily solar radiation and air temperature at 2 m height were obtained from the project solar data (SoDa) server (available online at http://www.soda.is.com/eng/index.html; Wald et al. 2002). The downward solar radiation data were found from the daily values of the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis. The server makes online queries to the archive (from 1958 onward) of the NCEP–NCAR database for radiation parameters, air temperature, and precipitation. The server is mirrored by the Institut Pierre-Simon Laplace in France. The hourly and daily net longwave irradiance was calculated using the Allen et al. (1998) procedure. The heat flux into the ground ($G$) varies during the day, and it depends on many factors such as leaf area index and soil moisture. However, it is strongly correlated with net radiation, and it is often assumed to be 30% of the net radiation (e.g., Santanello and Friedl 2003). Hence, we employed the same assumption in this study for simplicity. In open range areas, $K_w$ was assumed to be 0.5 to account for the low soil moisture availability that resulted from limited rainfall amounts at the sites that have open range areas. For all other vegetation types, $K_w$ was set equal to 1.0 throughout the study because green urban and irrigated olives areas were irrigated and because it was a wet year in Irbid. Following the recommendations and using the equations of Allen et al. (1998), the vapor pressure needed for ET$_0$ [Eq. (18)] was found by assuming the dewpoint temperature for
each day was equal to the minimum temperature for that day.

c. Validation data

A validation study was undertaken using data from the Agricultural Research Station of the University of Jordan (ARSUJ) in the central Jordan Valley at 32°10′N latitude and 35°37′E longitude at an altitude of ~230 m (below mean sea level). The station has a warm climate in winter, a minimum temperature of 8.5°C in January, and a hot summer, a maximum temperature of 40.4°C in July. The yearly average maximum and minimum temperatures are 30.9°C and 18.5°C, respectively, whereas the yearly mean temperature is 24.7°C. The experiment site has been selected in an alfalfa field where an automatic weather station (Campbell Scientific) has been installed. The crop is irrigated with a sprinkler irrigation system 2–3 times a week and planted on a sandy loam soil with good internal drainage. Data collected by the weather station include hourly and daily net solar radiation measured by a NRLITE-L net radiometer (Kipp and Zonen), hourly and daily wind speed at 2 m measured using a RM Young wind sentry 03101-5 system (Campbell Scientific), and air temperature and humidity measured at a height of 2 m using a shielded and aspirated REBS THP. ALARM was applied with satellite data obtained as daily and hourly solar radiation, air temperature, and wind speed. The American Society of Civil Engineers’ (ACSE 2005) alfalfa equation was used for this site. ASCE used daily solar radiation, air temperature, wind speed, and relative humidity. Net radiation was obtained using Allen et al.’s (1998) recommendations.

ALARM was applied with satellite data obtained as discussed in section 2b. The methods of ACSE (2005) were substituted for those of FAO-56. The equations and procedures of the two methods are very similar, except ACSE (2005) has a calibrated equation for potential evaporation from an alfalfa crop. While the behavior of the FAO-56 and ACSE (2005) methods should be very similar, it is possible that FAO-56 would have greater errors with the validation data than ACSE (2005) because of the calibration of the latter with an alfalfa crop.

The soil water content was monitored with TRIME tube access probe (P3, IMKO Micromodultechnik GmbH). The access tubes were 1 m in length and five probes were installed in the field. The measurements of the volumetric soil water content with the TRIME probe at depths of 0–20, 20–40, 40–60, and 60–80 cm were conducted manually once a day in the morning from March to October 2006. A water balance equation was used to calculate the measured ET using the soil moisture readings as

\[ ET_m = I - D - \Delta W, \]  

where \( ET_m \) is measured ET (mm day\(^{-1}\)), \( I \) is irrigation (mm day\(^{-1}\)), \( D \) is vertical drainage (mm day\(^{-1}\)), and \( \Delta W \) is the change in soil water (mm day\(^{-1}\)). Only \( ET_m \) for days of no irrigation (\( I = 0 \)) and zero \( D (D = 0) \) were used in this study to minimize the errors of \( ET_m \). Because of the limited number of usable satellite overpasses needed for use with ALARM, a total of 12 days of data were available for the validation study.

4. Validation study results and discussion

Results from the validation study in the ARSUJ alfalfa field are shown in Figs. 2–4. Because the field was irrigated, ET rates ranged from about 6 to about 10 mm day\(^{-1}\). For this range, the root-mean-square error (RMSE) for ALARM (as compared to the TRIME probe reference values, hereafter “measured” values) was 0.87 mm day\(^{-1}\), and the coefficient of determination \( r^2 \) was 0.36. The RMSE for ACSE (2005) was 1.25 mm day\(^{-1}\) and \( r^2 = 0.06 \).

While the RMSE of both methods with the measured values was relatively good, the lack of data from non-irrigated fields makes it difficult to estimate the uncertainty of the two estimates at lower evaporation rates. Errors in the ACSE (2005) and FAO-56 methods for moisture-stressed sites are likely to be dominated by errors in the water stress coefficient. These errors are likely to be quite large when taken as a percent error of the daily ET but relatively small in actual magnitude during very dry conditions. Assuming independent random errors equal to 1.25 mm day\(^{-1}\) in successive measurements, two daily FAO-56 measurements that differ by less than \( (1.25^2 + 1.25^2)^{1/2} \) mm day\(^{-1} \) = 1.76 mm day\(^{-1} \) are not far enough apart to rule out random variability.

Error magnitudes in ALARM are unlikely to vary greatly with the magnitude of the ET rate. Theoretically, EF varies with \( \Delta_T \), which is a ratio of two temperature differences (14). Absolute errors in both \( T_i \) and \( T_{\text{max}} \) are likely to be largest under conditions of high available energy, but \( T_{\text{max}} - T_i \) and \( T_{\text{max}} - T_o \) are likely to be large under these conditions; therefore, the relative errors are likely to be similar for a wide range of conditions. Experimentally, Suleiman and Crago (2004) applied the dimensionless temperature approach (without ALARM) to grasslands under stressed and un-stressed conditions and found similar scatter of estimated-to-measured ET under high (up to about 5.5 mm day\(^{-1}\)) and low (down to about 1.5 mm day\(^{-1}\)) daily ET.
rates. Thus, assuming random and independent errors, it seems reasonable to assume that two daily ALARM measurements of ET that differ by less than \((0.87^2 + 0.87^2)^{1/2}\) mm day\(^{-1}\) are not far enough apart to rule out random variability. Finally, if ET\(_{aa}\) and ET\(_{fa}\) estimates are different by less than \((1.25^2 + 0.87^2)^{1/2}\) mm day\(^{-1}\), they are not far enough apart to rule out random variability.

5. Results from the test sites

Air and MODIS radiometric surface temperatures in April 2002 for the different sites are shown in Fig. 5. The two highest radiometric surface temperatures and the highest differences between radiometric surface and air temperatures were seen in Rwaished and Safawi. The difference between radiometric surface

[Fig. 2. Measured, ALARM, and ASCE evapotranspiration measurements vs DOY at the validation site.]

[Fig. 3. ALARM vs measured evapotranspiration values at the validation site.]
and air temperatures was lowest in Irbid, where it was negative in some instances. Dimensionless temperature, $\Delta_T$, was greater than 0 for all the days for all the locations except Irbid, where it had six instances of negative $\Delta_T$, two of which were $<-0.50$ (Fig. 6). The influence of warm advection at Irbid will be discussed later. The positive $\Delta_T$ values ranged from 0.09 for Irbid to about 0.80 for Rwaished and Safawi.

The maximum ALARM actual ET, $ET_{aa}$, ranged from 2.4 mm d$^{-1}$ in Safawi to 6.5 mm day$^{-1}$ in Irbid. The maximum FAO-56 actual ET, $ET_{fa}$, ranged from 1.7 mm day$^{-1}$ in Mafrak to 7.1 mm day$^{-1}$ in Irbid (Table 4). The minimum $ET_{aa}$ ranged from 0.7 mm day$^{-1}$ in Safawi to 1.5 mm day$^{-1}$ in Mafrak, whereas the minimum $ET_{fa}$ ranged from 0.7 mm day$^{-1}$ in Aqaba to 3.7 mm day$^{-1}$ in Irbid. The average of $ET_{aa}$ and $ET_{fa}$ were almost identical in Amman; within 0.7 mm day$^{-1}$ in Rwaished, Safawi, Mafrak, and Aqaba; and greater than 1 mm day$^{-1}$ in Irbid (Table 4). These results indicate that there is generally good agreement between minimum, maximum, and average $ET_{aa}$ and $ET_{fa}$ at all sites except for Irbid, where the minimum, and consequently the average, $ET_{aa}$ and $ET_{fa}$ were different.

At three of the sites (Rwaished, Safawi, and Amman), $ET_{aa}$ and $ET_{fa}$ were similar throughout April (Fig. 7). At the Amman site, $ET_{aa}$ appeared to fluctuate from one day to another throughout April. However, $ET_{aa}$ and $ET_{fa}$ from days of year (DOY) 95–104 fluctuated over a wider range than they did from DOY 108–120. In the case of $ET_{fa}$, the fluctuations were smaller than 1.76 mm day$^{-1}$ so random variations cannot be ruled out. The $ET_{aa}$ fluctuations are on the order of 2 mm day$^{-1}$, which suggests they cannot be explained by random variations. One possible explanation is that soil moisture was reduced after DOY 104 and was unable to supply sufficient water to allow for high ET after this date. The drop in $ET_{aa}$ near DOY 105 at Aqaba, which is also greater than 1.2 mm day$^{-1}$, might have a similar explanation.

On most days, Mafrak and Aqaba had insignificant differences in $ET_{aa}$ and $ET_{fa}$. However, on some days there were larger differences. $ET_{aa}$ was consistently a little higher than $ET_{fa}$ in Mafrak and on most days in Aqaba. In Irbid, $ET_{aa}$ and $ET_{fa}$ were close on many days, whereas $ET_{aa}$ was lower than $ET_{fa}$ on several days, indicating that water stress may have taken place on these days. Because of its use of radiometric surface temperatures, $ET_{aa}$ is expected to respond automatically to water stress, whereas $ET_{fa}$ can only respond by changing $K_s$. On the other hand, on four days $ET_{aa}$ was greater than $ET_{fa}$, indicating that warm advection was present. For five days, $ET_{fa}$ was greater than 6 mm day$^{-1}$ and for DOY 111, $ET_{aa}$ was more than the reference ET ($ET_0$).

The $ET_{aa}$ and $ET_{fa}$ were much lower than $ET_0$ for all the sites except for Irbid. Generally, a good agreement was observed in the arid and semiarid sites, which were used as open rangeland, and in Amman, where the irrigated area was 21%. On the other hand, little agreement was observed in Mafrak, where 70% of the veg-
The site was denser as a result of the large fraction of protected rangeland, and in Irbid, where 75% of the site was cultivated, mainly with field crops (Table 2). For the sites of Safawi and Amman, the agreement between ETaa and ETfa was obvious, and both curves nearly coincided after DOY 105 (Fig. 7). A similar trend with less agreement was observed in Rwaished and Aqaba.

In general, ETaa had a well-defined linear relationship with ΔT, with some fluctuations resulting from the availability of energy (Rn – G) for ET (Fig. 8). This relationship, which is mathematically described in Eq. (16), demonstrated that the dimensionless temperature and Rn – G are responsible for actual ET determination. Although Rn – G is the main factor that influences potential ET, ΔT is a measure of the actual ET. Many points from the Irbid site deviated from the line because of the sensible heat advection, which resulted in higher actual ETaa at very low—even negative—ΔT.

The net radiation (Rn) ranged from about 10 to 16 MJ m⁻² day⁻¹ at the different sites (Fig. 9). The difference of Rn from one day to another depends on two factors. First, the daily clear-sky Rn increases systematically from 1 April to 30 April. Second, the actual Rn that is retained by the land surface depends on how clear (amount of clouds) the day is. The more cloudy

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**Fig. 5.** Air (Tair) and radiometric surface temperature (T) on different DOY during April 2005 for the six sites.
the day is, the less the actual \( R_n \) that is retained by the land surface. The maximum EF (\( \text{ET}_{aa}/R_n \)) was less than 0.6 for all the sites except for Irbid, where it was 1.3. In Irbid, EF was greater than 0.6 on 11 days and was greater than 1 on 3 days. The EF in Irbid was also frequently greater than at other locations, demonstrating the effect of the sensible heat advection on ET. A trend of increasing EF with \( R_n \) was observed for all sites with more scattered patterns for Irbid and Mafraq.

6. Discussion
Because \( \text{ET}_{aa} \) and \( \text{ET}_{fa} \) are both estimates of actual ET, neither is considered to be the correct, or refer-

<table>
<thead>
<tr>
<th>Stations</th>
<th>( \text{ET}_{aa} )</th>
<th>( \text{ET}_{fa} )</th>
<th>( \Delta_T )</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>Min</td>
<td>Max</td>
<td>Avg</td>
</tr>
<tr>
<td>Rwaished</td>
<td>0.8</td>
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<td>1.6</td>
</tr>
<tr>
<td>Safawi</td>
<td>0.7</td>
<td>2.4</td>
<td>1.5</td>
</tr>
<tr>
<td>Mafraq</td>
<td>1.5</td>
<td>3.2</td>
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<tr>
<td>Irbid</td>
<td>1.2</td>
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<tr>
<td>Amman</td>
<td>0.9</td>
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<td>Aqaba</td>
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ence, value against which the other is statistically compared. Rather, the two predictions are used to verify the reasonableness of each other, and their differences are used to shed light on the underlying land surface processes.

Ideally, $\Delta_T$ ranges from 0 when ET is maximum ($ET = R_n - G$) to 1 when ET is minimum ($ET = 0$). However, negative $\Delta_T$ can be found when warm advection takes place, and the air temperature becomes higher than the aerodynamic surface temperature. The warm air in this case acts as another source of energy through downward (negative) sensible heat flux. A low value of $\Delta_T$ may suggest that a soil has sufficient moisture available to meet ET demands, whereas a high value of $\Delta_T$ would indicate limited soil moisture availability. This is why Irbid, which received a relatively high rainfall amount, had the lowest $\Delta_T$ values, and Rwaished and Safawi, which did not receive much rain, had the highest $\Delta_T$ values (Table 4). A continuous increase of $\Delta_T$ is a clear signal of decreasing water availability and increasing water stress. This coincided with the trend of herbaceous vegetation growth in Mafraq, where the green flush continued for a short period and quickly reduced the available soil moisture. Under these conditions, both soil water and vegetation type and density would likely affect the estimate of actual ET$_{fa}$ from the FAO-56 equation (through the coefficients $K_s$ and $K_c$, respectively). This was most clearly seen in Irbid and Mafraq, where ET estimates from ALARM and FAO-56 were closest in periods following rainfall (Fig. 7). Actual ET from both estimates was in relatively good agreement for the other sites that were less vegetated and were either urbanized or grazed rangelands.

**FIG. 7.** Daily actual ALARM (ET$_{aa}$), FAO-56 (ET$_{fa}$), and FAO-56 grass reference (ET$_0$) evapotranspiration and daily rainfall amount on DOY during April 2005 for the six sites.
Fluctuations of $\Delta_T$ from one day to another could also result from the variability of the net solar radiation from one day to another. This was an expected behavior in all sites because the 1-km spatial resolution included a mixture of land use/covers. The presence of a large increase and then a decrease in $\Delta_T$ during drying periods (e.g., Mafraq around DOY 103) could result from either a nonrecorded small rain event that in-

![Graph](image-url)

**Fig. 8.** Relationship between the ALARM daily actual evapotranspiration ($ET_a$) and the dimensionless temperature ($\Delta_T$) for the six sites.

![Graph](image-url)

**Fig. 9.** Relationship between the daily evaporative fraction (EF) and the daily net radiation ($R_n$) for the six sites.
increased $\Delta_T$ in one day and then in the next $\Delta_T$ came back to where it was, or that the high $\Delta_T$ was an outlier for some reasons. Nevertheless, a $\Delta_T$ value of about 0.35 to 0.40 suggests that potential ET is taking place, whereas a $\Delta_T$ value of about 0.75 to 0.80 implies that soil evaporation is not contributing significantly to ET because of a dry soil surface and transpiration is not meeting the evaporative demand. Any $\Delta_T$ that is significantly less than about 0.30 is likely a result of sensible heat advection because when $\Delta_T$ is about 0.30, ET seems to be close to potential ET.

The decreasing trends of actual ET may reflect dry-down cycles of soil moisture as a result of ET and deep percolation. The contribution of deep percolation to the loss of soil water cannot be assessed without monitoring or simulating the change of the soil moisture profile due to drainage (Suleiman and Ritchie 2004). The values of actual ET$_{aa}$ and its range seemed reasonable, although no site-specific calibration has been done. The trend of increasing ET$_{aa}$ from DOY 100 to DOY 104 in Aqaba could be attributed to the increase of air temperature during this period and the irrigation activities to the south of the site. The decrease of ET after that period was attributed to the loss of soil water from the surrounding areas.

Because of the warm advection and the high wetness conditions, the maximum and average actual ET were higher in Irbid than all the other sites (Table 4). The advection in Irbid is a well-known phenomenon that usually results in crop failure, particularly lentil crops (National Center for Agricultural Research and Extension 2002a). Comparing the daily air temperature records (National Center for Agricultural Research and Extension 2002b) with the long-term averages (Jordan Meteorological Department 2002) showed that the maximum air temperature was 7°C above the average during this period. During the periods of advection, the maximum air temperatures ranged from 29°C to 31°C, compared with an average value of less than 21°C. These results concurred with Rosenberg and Verma (1978) who found, in a study on irrigated alfalfa, that ET was greater than 10 mm day$^{-1}$ on one third of the study days. They attributed these high ET values to the contribution of sensible heat advection to ET. In this study, the effect of advection on ET estimates was observed by ALARM and MODIS data, whereas potential ET equations, such as Priestley–Taylor, are not capable of predicting these effects because they are developed for minimally advective conditions (e.g., Brutsaert 1982).

The interaction of soil water and climatic factors resulted in high ET values in Irbid. Suleiman and Ritchie (2003) demonstrated that soil evaporation could be more than 10 mm day$^{-1}$ when the saturated soil layers are close to the surface. This was possible in Irbid because the area had heavy clayey soils (Ministry of Agriculture 2008) known as vertisols. These soils are well-known for their high water holding capacity. Therefore, higher values of ET were obtained for this site because water from heavy rainfall was stored in the soil. However, the other sites were characterized by arid soils that received low rainfall amounts and, subsequently, lower ET values were noticed at these sites. Therefore, the impact of soil water availability and advection on the estimation of actual ET would suggest that ALARM might be able to provide a more accurate estimate of ET than the FAO-56 approach, which assumed no water stress (Allen et al. 1998).

It should be kept in mind that the estimation of actual ET from ALARM and the FAO-56 approach may produce uncertainties. The implementation of the constant evaporative fraction assumption would yield accurate daily ET estimates for cloud-free days, as indicated by Zhang and Lemeur (1995). However, variations in cloudiness during the day may produce some uncertainties in ET estimates because clouds at times other than the when the satellite passes may invalidate the constant EF assumption. This effect, that is, varying EF during midday hours, has been found from numerous observations to vary little (e.g., Shuttleworth et al. 1989; Gurney and Hsu 1990, Brutsaert and Chen 1996; Crago and Brutsaert 1996; Lhomme and Elguero 1999), and it has been found that it depends on the site and time of year (Kustas et al. 1993). In semiarid areas under a wide range of conditions, Kustas et al. (1993) found that the correlation coefficient between midday and daytime EF was rather high ($r = 0.92$). Estimating actual ET from the FAO-56 approach, on the other hand, would involve the use of an assumed value of $K_c$, which could result in some uncertainties, even under available water and irrigation (Jitan 2005), as indicated by the present ET estimates in Mafraq and Irbid.

7. Summary and conclusions

This research describes a case study of comparing the application of remote sensing and modeling for actual ET estimation with the well-established FAO-56 procedure. Although remote sensing cannot provide a direct measurement of ET (Venturini et al. 2004), the contribution of remotely sensed data is generally considered to be invaluable, particularly in a country like Jordan, where droughts are frequent and meteorological data are scarce. The implementation of ALARM to convert MODIS radiometric to aerodynamic surface temperature at a well-defined scalar roughness length and the use of ALARM along with the dimensionless
temperature procedure proved to be valuable in providing actual ET estimates of different ecological zones. The approach provides the actual ET without the need for soil moisture measurements and was able to detect ET characteristics at the different sites in Jordan. These estimates, in addition to remotely sensed indices, would provide a real-time monitoring system that could detect dry cycles and vegetation response to rainfall. Therefore, future efforts will focus on the validation, calibration, and application of MODIS data, ALARM, and dimensionless temperature to estimate actual ET in Jordan. Following that, a hydrologic tool will be developed to broadcast efficient, accurate daily actual ET maps. Also, the integration of MODIS data with near-real-time monitoring may be used to estimate actual crop ET to contribute to irrigation water management in the country. As part of the extension services, the climatic data needed for this purpose are becoming more available on daily, weekly, and monthly bases on the Web (available online at http://www.merimis.org/).

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