Estimation of Evapotranspiration Using SEBAL Algorithm and Landsat-8 Data—A Case Study: Tatra Mountains Region

Ayad Ali Faris Beg¹, Ahmed H. Al-Sulttani², Adrian Ochtyra³,⁴, Anna Jarocińska³ and Adriana Marcinkowska³

1. Department of Geography, College of Education, University of Mustansiriyah, Baghdad, Iraq
2. Department of Regional Planning, College of Physical Planning, University of Kufa, Najaf, Iraq
3. Department of Geoinformatics, Cartography and Remote Sensing, Faculty of Geography and Regional Studies, University of Warsaw, Warsaw, Poland
4. College of Inter-Faculty Individual Studies in Mathematics and Natural Sciences, University of Warsaw, Warsaw, Poland

Abstract: ET (Evapotranspiration) is one of the climate elements, which plays an important role in water balance, and effects on the ecosystem of any region. Therefore, many mathematical equations and algorithms have been found and designed to calculate and estimate values of evapotranspiration. Calculation methods are either based on data from meteorological stations or using other sources of data where the area is lacking from meteorological stations. Remote sensing data are one of the important sources and techniques to estimate many climate elements including evapotranspiration. The selected study area is located in Tatra Mountains on the borders between Poland and Slovakia. Tatra Mountains are the most valuable areas in Poland and Slovakia. The main objective of current study is to estimate the spatial variation of ET using SEBAL algorithm and Landsat-8 imagery. The analysis is carried out using Landsat-8 (OLI/TIRS) data, ASTER GDEM and reference weather parameters. Sixteen ERDAS models are prepared to calculate the various parameters related to solar radiation. The models are prepared to calculate the values of surface radiance, surface reflectance, surface albedo, NDVI, LAI, surface emissivity, surface temperature, net radiation, soil heat flux, sensible heat flux, latent heat flux, which are consequently used to calculate the hourly and daily evapotranspiration in study area. Results of pixel wise calculations show the values of surface temperature which are varied from 6.2 °C at mountain shadow areas to 34.6 °C at bare rocks and bare land area, while the spatial variation of ET at different land covers shows the hourly ET ranged from 0 to 0.72 mm/hr, while the daily ET varied from 0.0 to 17.0 mm/day. Results show clear relation between land use/land cover and solar radiation parameters and impact of vegetation cover on the ET values in pixel wise domain.

Key words: Landsat-8, remote sensing, evapotranspiration, solar radiation, SEBAL, NDVI.

1. Introduction

With growing attention to climate elements and climatic changes and their effects on the environment and ecological systems, the interesting of evapotranspiration estimation is increased as well, especially with the emergence of various measuring techniques for assessment of climate elements. Recently, the remote sensing becomes one of the valuable data sources and analysis technique for implementing such studies. Accurate estimation of evapotranspiration is considered as the key factor in water management [1]. ET (Evapotranspiration) is a climate element being an important factor for energy, hydrologic, carbon and nutrient cycles and a key item of the water balance in the soil-vegetation-atmosphere continuum [2]. The major advantage of applying remote sensing is the water consumed as ET can be derived directly without needing for quantifying other complex hydrological processes [3]. Merlin et al. [4] referred to the importance of ET in predicting soil water availability, flood forecasting, rainfall forecasting and projecting changes in occurring heat
waves and droughts.

Nowadays, many researches have been conducted to gather spatially distributed ET over large scales by using surface energy balance and remote sensing data. These techniques provide spatial information from the earth’s surface by measuring reflected and emitted electromagnetic radiation. The measurements of thermal infrared, infrared and visible bands of remote sensing data are the inputs for the parameterization of the energy balance components in the calculation of ET [5]. The SEBAL (Surface Energy Balance Algorithm for Land) is used for estimating various land surface parameters i.e., surface albedo, surface temperature, and energy balance parameters, from different spatial, spectral and radiometric resolutions of satellite data. Consequently, the consistency of ET estimation from different satellites by SEBAL needs to be certified [6]. Venturini et al. [7] and Courault et al. [8] categorized the methods used in surface flux i.e., empirical direct methods where remote sensing data are introduced directly in semi-empirical models to estimate ET, residual methods of the energy budget combining some empirical relationships and physical models. Most current operational models such as (SEBAL, and deterministic methods generally are based on more complex models such as SVAT (Soil-Vegetation-Atmosphere Transfer) models), which calculate the different components of the energy budget. Many methods and models are designed for estimating the evapotranspiration using remote sensing data. SEBAL is an ET estimation approach based on satellite images via computing a land surface energy balance. The model is developed by Bastiaanssen et al. [9] and has been validated at many locations around the world [10]. The model represents the most promising algorithm that needed minimum input data of ground based variables and it has been widely applied abroad the world accurately to estimate actual ET [1]. The surface energy balance provides through latent heat flux a direct assessment of actual ET [11]. SEBAL method uses multispectral remote sensing data associated with complementary meteorological data to estimate instantaneous surface energy balance components [12]. The overall accuracy of ET from SEBAL is around ±15%. Accuracy can be improved by space and time integration, while the seasonal differences ranged from 1% to 5% due to a reduction in the random error component [11]. Bastiaanssen et al. [11] proved the importance of SEBAL method in establishing a variety of measurements and assessments, including land use and water use for river basin planning, impact of water conservation, environmental impact assessment, hydrological modeling and more other application.

In Tatra Mountains region, the reliance on meteorological stations in collecting of evapotranspiration reading is impossible, because no local weather stations measure evapotranspiration, so the remote sensing data will be the most quick and trusted source to estimate evapotranspiration. The main objective of current study is to use Landsat-8 data and SEBAL method for estimating the distribution of evapotranspiration pattern in study area based on building several analysis models using ERDAS imagine software.

2. Scope of Study Area

Study area is located in the middle-east of Europe and in the south of Poland on the borders between Poland and Slovakia in Tatra mountains. The area is bounded by latitudes 49º 8’-49º 22’ N and longitudes 19º 38’-20º 16’ E, and extended over 1,280 square kilometers (Fig. 1). Tatras are the highest mountains and relatively young because they are from Alpine orogeny [13]. It can be divided into four parts: Western Tatra Mts., High Tatra Mts., Bielskie Tatra Mts. and Siwy Wierch massif [14]. The tectonic uplift movements of Tatras cause mixing igneous rocks, metamorphic rocks and sedimentary rocks. The High Tatras are constructed of igneous rocks, mainly granites. Western Tatras are mostly built of metamorphic rocks (gneisses and schists), igneous
The Bielskie Tatra Mts. and the Siwy Wierch massif are built of limestone [15, 16]. The analyzed region is characterized by vertical zones of climate (different temperatures and precipitation), soils and vegetation. In Tatras, it is possible to distinguish five climatic-vegetation belts: lower montane, upper montane, dwarf pine, alpine and subnival [17]. The lower montane (700-1,200 m a.s.l.) is covered by spruce-fir-beech forests; upper montane (1,200-1,550 m a.s.l.) is also covered mainly by spruce (*Picea abies*); dominating species are mainly the dwarf pine (*Pinus mugo*). Alpine belt (1,800-2,300 m a.s.l.) is covered by different grasses or dwarf shrubs, and in the highest belt-subnival (2,300-2,655 m a.s.l.) are bryophytes, lichens and bare rocks.

**3. Method of Analysis**

Spatial variation of ET depends on land use/land cover in any area. To estimate ET values several solar radiation parameters need to be calculated. The main source of data is represented by Landsat-8 data OLI/TIRS with Path 188/Row 26, acquired on 08-SEP-2013, and one arc-second resolution ASTER GDEM with some reference weather parameters nearby the area are used. Land use/land cover classes in study area are classified using SAM (spectral angle mapper) classification method with spectral angle of 0.3 using ENVI v.5.3 software.

To calculate the ET using SEBAL method, the required mathematical equations for building the algorithms of processing and analysis models are collected. Sixteen spatial analysis models are prepared using ERDAS imagine software i.e. atmospheric correction models for calculating the reflectance, surface albedo, emissivity, surface temperature, and elements of solar radiations balance, including net radiations, soil heat flux, sensible heat flux, latent heat flux, and finally calculating the instantaneous and daily evapotranspiration values.

**3.1 Atmospheric Correction of Landsat-8 Data**

Atmospheric correction for solar radiations is important in remote sensing analysis; its necessity depends on the objectives of the analysis. In general, land cover identification exercises that are based on single-date images do not require atmospheric correction, as pixels are being compared with other pixels within an image in terms of similarity [18]. Estimation of the ground target reflectance starts with converting the pixel value to radiance [18]. Accordingly, Landsat-8 OLI/TIRS bands data can be converted to TOA spectral radiance using the radiance
rescaling factors provided in the metadata of images [19]:

\[ L_\lambda = M L * Q_{cal} + A L \]  

Where \( L_\lambda \) = TOA spectral radiance (Watts/(m²*srad*μm)); \( ML = \) Band-specific multiplicative rescaling factor from the metadata (RADIANCE_MULT_BAND_x, where \( x \) is the band number); \( AL = \) Band-specific additive rescaling factor from the metadata (RADIANCE_ADD_BAND_x; \( x \) is the band number); \( Q_{cal} = \) Quantized and calibrated standard product pixel values (DN).

To compute the top of planetary reflectance based on spectral radiance at the sensor aperture is by following the equations given by Refs. [20, 21]:

\[ \tau_\lambda = \frac{\pi \times L_\lambda}{ESUN_\lambda \times \cos \theta \times d_r} \]  

\( \rho_\lambda = \) TOA planetary reflectance
Where \( ESUN_\lambda \) = Exoatmospheric spectral solar irradiance on a surface perpendicular to the sun’s ray (Wxm⁻²xμm⁻¹) as given by Ref. [19] (Table 1).

\[ \cos \theta = \text{solar zenith} = (90 - \text{solar elevation}) \]  
\[ \theta \] = Is in decimal degree need to be converted into radians as follows:

\[ \theta \text{ in radians} = \frac{x}{180} \times (\theta \text{ in decimal degrees}) \]  

\( d_r = \) Earth-Sun distance in astronomical.

### 3.2 Calculation of Surface Albedo

According to the equations mentioned in Refs. [21, 23, 24], the values of albedo for the top of atmosphere are calculated as:

\[ \alpha_{\text{TOA}} = \frac{\sum \omega_\lambda \times \rho_\lambda}{\sum \omega_\lambda} \]  

\( \omega_\lambda = \frac{ESUN_\lambda}{\sum \omega_\lambda} \)

Where \( \omega_\lambda = \) constant value of weighting coefficient.

\[ \alpha_{\text{TOA}} = \frac{\sum_i^n \text{Ref}_i \times \text{Si} \times \Delta \lambda_i}{\sum \text{Si} \times \Delta \lambda_i} \]  

Where \( Si = \) Solar irradiance constant, \( \text{Ref}_i = \) Reflectance of band \( i \) and \( \Delta \lambda_i = \) Band range.

So, the equation will be as follows:

\[ \alpha_{\text{TOA}} = 0.356 \times \text{Ref}_3 + 0.326 \times \text{Ref}_5 + 0.138 \times \text{Ref}_4 + 0.084 \times \text{Ref}_5 + 0.056 \times \text{Ref}_6 + 0.041 \times \text{Ref}_7 \]  

Where \( \alpha_{\text{TOA}} \) is the weighted albedo and \( rfl_{2,3,...7} \) are the planetary reflectance values for each band.

The weighted albedo is converted to surface albedo based on the following equation [21, 24-26]:

\[ \alpha = \frac{\alpha_{\text{TOA}} - \alpha_{\text{path–radiance}}}{\tau_{sw}^2} \]  

Where \( \alpha \) is surface albedo; \( \alpha_{\text{path–radiance}} \) is the incoming shortwave radiation flux reflected back to the sensor (ranged from 0.025 to 0.04), in SEBAL the value of 0.03 is used.

\( \tau_{sw} \) is the atmospheric transmissivity, it is calculated as:

\[ \tau_{sw} = 0.75 + 2 \times 10^{-5} \times z \]  

Where \( z \) is an elevation of area where is defined by ASTER GDEM Data.

### 3.3 Calculation of Emissivity and Land Surface Temperature

To calculate the surface temperature, first the black body temperature at satellite must be calculated [20, 27]:

\[ T_b = \frac{K_2}{\ln(K_1 \times L_\lambda + 1)} \]  

Where \( T_b \) is black body temperature at satellite in Kelvins, \( K_1 \) and \( K_2 \) are calibration constants for Landsat-8 OLI/TIRS thermal band-10, their values are \( K_1 = 774.89 \ W \text{m}^{-2} \text{ster}^{-1} \mu \text{m}^{-1} \), and \( K_2 = 1321.08 \) in Kelvin, \( L_\lambda \) is spectral radiance in \( W \text{m}^{-2} \text{ster}^{-1} \mu \text{m}^{-1} \), and is calculated by the following equation [27]:

### Table 1 Exoatmospheric spectral solar irradiance (ESUN) of Landsat-8 OLI bands [22].

| Band | 2 | 3 | 4 | 5 | 6 | 7 | 9 |
|------|---|---|---|---|---|---|---|
| TOA-ESUN | 2,067 | 1,893 | 1,603 | 972.6 | 245.0 | 79.72 | 399.7 |
To compute the land surface temperature, the black body temperature is corrected with respect to the surface emissivity \( \varepsilon \) values. The surface emissivity is a factor that describes the efficiency of an object radiates energy in comparing with black body [28]. The values of emissivity are estimated from NDVI and LAI as follows:

\[
\varepsilon = 1.009 + 0.047 \times \ln(NDVI) \tag{13}
\]

Where \( NDVI > 0 \), otherwise, emissivity is assumed to be zero (e.g. water) [21, 26].

\[
\varepsilon = 0.97 + 0.0033 \times LAI \tag{14}
\]

For \( LAI < 3.0, \varepsilon = 0.98 \) when \( LAI \geq 3 \);
For water, \( NDVI < 0 \) and surface albedo < 0.47, \( \varepsilon = 0.985 \) [25].

NDVI (Normalized Difference Vegetation Index), and LAI (Leaf Area Index), are calculated by visible and near infrared bands [25, 28].

For Landsat-8,

\[
NDVI = \frac{(B5-B4)}{(B5+B4)} \tag{15}
\]

To calculate the values of LAI, and SAVI (soil adjusted vegetation index) need to be calculated first.

\[
SAVI = \frac{(1+L) \times (TIR - Re\,d)}{(L + TIR + Re\,d)} \tag{16}
\]

Where \( L \) is a constant, which depends on the area properties, \( L \approx 0.5 \) is used, then the LAI is calculated from the empirical equation [25]:

\[
LAI = -(1/a2) \times \ln(a0 - SAVI/a1) \tag{17}
\]

Where \( a0, a1, a2 \) are 0.84, 0.65 and 0.6 respectively (these values are used to be compatible with study area conditions), then the corrected land surface temperature \( T_s \) is calculated [27]:

\[
T_s = \frac{T_b}{1 + (\lambda \times T_b / \gamma) \times \ln \varepsilon} \tag{18}
\]

Where \( \lambda \) is the average of limiting wavelengths of band 10 of Landsat8-TIRS (\( \lambda = 10.895 \mu m \))

\[
\gamma = h \times c / a \ (0.01438 \text{ m.K}) \tag{19}
\]

\[
a = \text{Boltzmann constant} \ (1.38 \times 10^{-23} \text{ J.K})
\]

\[
h = \text{Plank’s constant} \ (6.626 \times 10^{-34} \text{ J.S})
\]

\[
c = \text{velocity of light} \ (2.998 \times 10^8 \text{ m/s})
\]

3.4 Calculation of Solar Radiations Elements and ET

Estimation of ET by SEBAL is a model to calculate ET from the remote sensing images. It depends on the thermo-dynamic equilibrium between turbulent transport process in the atmosphere and laminar process in the sub surface. This process called the land surface energy balance [9], also Ref. [5] refers to the energy exchange which governs the ET process at the vegetation surface can be expressed mathematically as:

\[
R_n = G + H + \lambda ET \tag{20}
\]

Where \( R_n \) is the net radiation in Watt/m². \( G \) is the soil heat flux in Watt/m². \( H \) is the sensible heat flux in Watt/m². \( \lambda ET \) is the latent heat flux (\( \lambda \) is the latent heat of vaporization, and \( ET \) is evapotranspiration), which represents the heat loss from the surface due to \( ET \) and calculated as a “residual” of surface energy balance [9, 29].

Net radiation is the actual radiant energy available at the surface. It is computed by subtracting all outgoing radiant fluxes from all incoming radiant fluxes [25, 30]:

\[
R_n = (1 - \alpha) \times R_s \downarrow + RL \downarrow - RL \uparrow - (1 - \varepsilon) \times RL \downarrow \tag{21}
\]

Where \( R_s \downarrow \) incoming shortwave radiation (Watt/m²), \( \alpha \) is the broadband surface albedo (dimensionless), \( RL \downarrow \) is the incoming longwave radiation (Watt/m²), \( RL \uparrow \) is the outgoing longwave radiation (Watt/m²) and \( \varepsilon \) is the surface emissivity (dimensionless).

The incoming shortwave radiation \( (R_s \downarrow) \) is computed from the available climatic parameters such as sunshine hours, relative humidity, maximum and minimum temperature cloud cover and geographic location [25, 31].

\[
R_s \downarrow = G_{Sc} \times \cos \theta \times d_i \times \tau_{sw} \tag{22}
\]

Where \( G_{Sc} \) is solar constant (1,367 W/m²), \( \theta \) is the
solar incidence angle $\theta = 90 - \phi$, where $\phi$ = sun elevation angle, given in the header data file of the LANDSAT_8 imagery, $d_1$ = Earth-Sun distance in astronomical, and $\tau_{sw}$ is the atmospheric transmissivity.

The outgoing longwave radiation ($RL_{\uparrow}$), is computed at each pixel using Stefan-Boltzmann equation as Refs. [25, 26]:

$$RL_{\uparrow} = \varepsilon \times \sigma \times T_s^4$$

Where $\varepsilon$ is the surface emissivity, $\sigma$ is Stefan-Boltzmann constant ($5.67 \times 10^8 \text{ W/m}^2/\text{K}^4$) and $T_s$ is surface temperature in K.

To compute the incoming longwave radiation ($RL_{\downarrow}$), the hot and cold pixels must be selected as anchor pixels. Cold pixel is selected from the area of High density of vegetation cover with LAI > 3 from irrigated crops, at which the surface temperature and near surface air temperature are assumed to be similar.

$$H = Rn - G - 1.05 \times ETr$$

Where, $ETr$ is the rate of ET from reference.

Sensible Heat Flux (H) is the rate of heat loss to the air by convection and conduction, due to temperature gradients [32, 33].

$$H = \rho \times C_p \times \frac{dT}{r_{ab}}$$

Where $H$ is sensible heat flux, $\rho$ is air density (kg/m³), $C_p$ is air specific heat (1.004 J/kg/K), $dT$ is near surface temperature differences in K, and $r_{ab}$ is aerodynamic resistance to heat transport (m/s). To get the stability in $r_{ab}$ value, the two anchors “hot” and “cold” pixels are used at two heights $z_1$ and $z_2$, to facilitate the computation of near surface temperature differences. Near surface temperature difference is calculated from the linear relationship between $dT$ and $T_s$ [10, 25].

$$dt = aT_s + b$$

Where $a$ and $b$ are the correlation coefficients which are calculated iteratively from the data of hot and cold anchor pixels using xls-spreadsheet prepared for that purpose.

The aerodynamic resistance to heat flux $r_{ab}$ is computed for neutral stability as:

$$r_{ab} = \frac{z_1}{u \times k}$$

Where $z_1$ and $z_2$ are heights in meter above the zero plane displacement of vegetation. The used values are 2.0 m, 0.1 m respectively. $u \ast$ is the friction velocity (m/s), $k$ is Von Katman’s constant (0.41).

$$u = k \times u \ast \times \ln(z_2 / z_{om})$$

Where $u_\ast$ is the wind speed (m/s) at height $z_2$ [33].

The momentum roughness length for each pixel is calculated as a function of LAI [25]:

$$z_{om} = 0.018 \times LAI$$

The atmospheric conditions of stability have a large
effect on the aerodynamic resistance ($r_{ah}$). The stability conditions are defined by using the Monin-Obukhov length ($L$), and computed as follows:

$$L = -\frac{\rho \times C_p \times u^3 T_s}{k \times g \times H}$$  \hspace{0.5cm} (32)

$\rho$, $C_p$, $T_s$, $k$ and $H$ are defined previously, $g$ is the gravitational constant (9.81 m/s$^2$).

The stability correlations for momentum and heat transport and $h$ are computed iteratively depending on the atmospheric conditions by using xls-spread sheet designed based on equations mentioned by Waters et al. [25]. The values of cold and hot pixels are adjusted to their elevations and calculated iteratively and then the last iteration values of $a$, $b$ and $r_{ah}$ are used in computing the sensible heat flux (H).

Latent heat flux is the rate of latent heat loss from the surface due to evapotranspiration [25].

$$\lambda \times ET = R_n - G - H$$  \hspace{0.5cm} (33)

$$ET_{inst} = 3600 \times \frac{\lambda \times ET}{\lambda}$$  \hspace{0.5cm} (34)

Where $\lambda$ is the latent heat of vaporization in J/kg.

$$\lambda = 2.501 - (T_a - 273) \times 0.002361$$  \hspace{0.5cm} (35)

Where, $T_a$ is the mean air temperature (K).

Then the reference ET fraction (ETf) computed as follows:

$$ET_f = ET_{inst} / ETr$$  \hspace{0.5cm} (36)

Where, $ETr$ is the reference $ET$ at the time of the image.

Daily values of $ET_{24}$ are computed by assuming that, the computed instantaneous $ET_{24}$ is the same as the 24-hour average. Therefore, the $ET_{24}$ (mm/day) can be computed as:

$$ET_{24} = ET_f \times ET_{r-24}$$  \hspace{0.5cm} (37)

Where, $ET_{r-24}$ is the cumulative 24-hour $ETr$ for the day of the image [25].

$$ET_{r-24} = \sum_{h}^{24} ET_{r-h}$$  \hspace{0.5cm} (38)

3.5 Building of Spatial Modeling

To achieve the requirements of SEBAL method for estimating the evapotranspiration on pixel wise, all the mathematical equations are used in building the algorithms of sixteen analysis models. The models are prepared by using ERDAS 2014 software (Fig. 2).

4. Results and Discussion

To outline pattern and behaviour of surface landsurface interaction with solar radiation elements, land use/land cover classification is carried out. The classification results of land use/land cover in the Tatra mountain region show seven classes (Fig. 3). The majority of the land cover which is controlling the solar radiation and ET regime is represented by coniferous forests with coverage of 51.6% of the total area, followed by grasslands with fields of vegetation covers about 34.9%, then the rock outcrops with 8.0%, and the remains with 5.5% including built up mixed with rocky area, hill shadows and water bodies.

Moreover, the analysis results (maps, statistics and graphs) show the spatial variation of solar radiation elements (Fig. 4) and evaptranspiration of pixel wise (Fig. 5). To define the distribution and dispersion amounts in solar radiations, ET and the causes of the variations, measuring of descriptive statistic parameters are carried out as shown in Tables 2 and 3 and Fig. 6. The spatial variation of the main solar radiation elements is analysed. The principle statistic measurements i.e., minimum, maximum, range, mean and standard deviation values are used to give indications about the characteristics of land cover classes. Mean values are the most important statistic parameters, which reflect the characteristics of most pixels that are controlling the solar radiations behaviour within each class of landcover. The results of surface albedo show the bare lands and rock outcrops are having the highest albedo with average values of 0.18 and 0.17 respectively, this because of the low absorption and scattering from such land cover types, while the forests, grasses and water bodies which are recognised by high absorption properties are having low average albedo values from 0.05 to 0.08.
Estimation of Evapotranspiration Using SEBAL Algorithm and Landsat-8 Data—A Case Study: Tatra Mountains Region

Fig. 2  Calculations of solar radiation parameters and evapotranspiration from Landsat-8.

Fig. 3  Land use/land cover in Tatra region.
Dealing with NDVI and LAI, the biomass of vegetation covers have been represented by forests, grasslands and vegetation fields can be recognized by higher average values of NDVI and LAI, while the rocks and water bodies give low to negative values. The values of emissivity for all the land cover classes are relatively close. However, the bare lands, built-up and mixed with rocky areas show the minimum range values of emissivity about 0.006 and 0.013 respectively, which are reflecting the homogeneity of the landcovers and low inclusions pixels of other classes. The land surface temperature shows high variations in the mean temperature values ranged from 14 °C in hills shadow area to 28.5 °C in bare lands. The maximum temperature values in the bare land area are 34.6 °C, while the minimum values in the hills shadows on the Tatras mountains about 6.2 °C. About the solar radiation balance which is summarised by the distribution pattern of instantaneous and daily evapotranspiration values in Table 3 and Figs. 5 and 6. Normally, the minimum values of instantaneous evapotranspiration in the built-up area and rock outcrops
and the maximum ET come from the shadow area in the top of the Tatras mountains which reach up to 0.721 mm per hour followed by grasslands and forests area. The grassland, rock outcrops and built-up area show high ranges of ET, that due to the presence of pixels of different landcovers within these classes. The results of the mean values of ET in the shadow area and forests give the highest values of 0.504 and 0.43 mm per hour respectively. The lowest mean values of ET come from the bare land area. The shadow area and forests are the main landcovers which are controlling the ET pattern in the study area. In the same sequence, the pattern of daily ET comes with higher average values at shadows and forest areas with 12.09 and 10.25 mm per day, respectively, bare lands show the lowest ET values with 4.48 mm per day only.
Table 2  Statistics of emissivity, albedo, NDVI, LAI and surface temperature (°C) for different land use/land cover in Tatra region.

| Land use/Land cover     | ID | %  | Area (km²) | S-Albedo | NDVI |       |       |       |       |       | Surface temperature (°C) |
|-------------------------|----|----|------------|----------|------|-------|-------|-------|-------|-------|--------------------------|
|                         |    |    |            |          |      | Min   | Max   | Range | Mean  | Std   | Min   | Max   | Range | Mean  | Std   |
| Rock outcrops           | 1  | 8.0| 102.0      | 0.038    | 0.48  | 0.44  | 0.17  | 0.05  | -0.056| 0.537 | 0.593 | 0.203 | 0.092 |
| Grass land with field of vegetation | 2  | 34.9| 446.8      | 0.033    | 0.44  | 0.40  | 0.13  | 0.03  | -0.103| 0.602 | 0.705 | 0.361 | 0.086 |
| Forests                | 3  | 51.6| 659.8      | 0.037    | 0.23  | 0.19  | 0.08  | 0.02  | -0.061| 0.567 | 0.628 | 0.279 | 0.076 |
| Bare lands             | 4  | 0.2| 2.7        | 0.080    | 0.28  | 0.20  | 0.18  | 0.02  | 0.063 | 0.483 | 0.420 | 0.204 | 0.062 |
| Hills shadows          | 5  | 2.4| 31.3       | 0.036    | 0.13  | 0.10  | 0.05  | 0.01  | -0.061| 0.370 | 0.431 | 0.051 | 0.064 |
| Water bodies           | 6  | 0.1| 1.0        | 0.042    | 0.06  | 0.02  | 0.05  | 0.00  | -0.062| 0.108 | 0.171 | -0.029| 0.009 |
| Built-up mixed with rocky area | 7  | 2.8| 36.2       | 0.049    | 1.04  | 0.99  | 0.15  | 0.03  | -0.052| 0.582 | 0.634 | 0.281 | 0.096 |

| ID | Min LAI | Max LAI | Range LAI | Mean LAI | Std LAI | Min Nb-emissivity | Max Nb-emissivity | Range Nb-emissivity | Mean Nb-emissivity | Std Nb-emissivity | Min surface temperature | Max surface temperature | Range surface temperature | Mean surface temperature | Std surface temperature |
|----|---------|---------|-----------|----------|--------|------------------|------------------|---------------------|---------------------|-------------------|------------------------|------------------------|--------------------------|--------------------------|-------------------------|
| 1  | -0.304  | 2.654   | 2.958     | 0.363    | 0.363  | 0.969            | 0.990            | 0.021               | 0.971               | 0.001             | 8.3                    | 34.2                   | 25.9                     | 23.2                    | 3.8                     |
| 2  | -0.304  | 4.177   | 4.481     | 1.130    | 0.363  | 0.969            | 0.990            | 0.021               | 0.974               | 0.002             | 8.1                    | 34.3                   | 26.2                     | 23.7                    | 3.3                     |
| 3  | -0.304  | 3.187   | 3.491     | 0.677    | 0.363  | 0.969            | 0.990            | 0.021               | 0.972               | 0.001             | 9.9                    | 30.6                   | 20.7                     | 20.5                    | 1.8                     |
| 4  | -0.133  | 1.995   | 2.127     | 0.345    | 0.363  | 0.969            | 0.990            | 0.021               | 0.971               | 0.001             | 22.5                   | 34.6                   | 12.1                     | 28.5                    | 2.2                     |
| 5  | -0.304  | 1.112   | 1.416     | 0.149    | 0.363  | 0.969            | 0.990            | 0.021               | 0.975               | 0.009             | 6.2                    | 22.4                   | 16.2                     | 14.0                    | 3.0                     |
| 6  | -0.304  | 0.002   | 0.306     | 0.303    | 0.363  | 0.969            | 0.990            | 0.021               | 0.981               | 0.010             | 12.9                   | 20.1                   | 7.2                      | 15.7                    | 1.4                     |
| 7  | -0.304  | 3.539   | 3.842     | 0.709    | 0.363  | 0.969            | 0.982            | 0.021               | 0.972               | 0.002             | 13.4                   | 31.8                   | 18.4                     | 25.9                    | 1.2                     |
### Table 3  Statistics of solar radiation elements and ET for different land use/land cover in Tatra region.

| ID | Net radiation | Soil heat flux | Sensible heat flux |
|----|---------------|----------------|--------------------|
|    | Min | Max | Range | Mean | Std | Min | Max | Range | Mean | Std | Min | Max | Range | Mean | Std |
| 1  | 220.4 | 611.5 | 391.1 | 446.1 | 48.3 | 56.8 | 305.7 | 249.0 | 110.7 | 15.6 | 44.3 | 190.2 | 145.9 | 128.2 | 21.5 |
| 2  | 232.3 | 610.1 | 377.8 | 473.8 | 35.4 | 59.0 | 305.0 | 246.1 | 119.4 | 12.4 | 43.5 | 190.9 | 147.4 | 131.5 | 18.4 |
| 3  | 403.0 | 600.9 | 197.9 | 526.2 | 21.4 | 69.2 | 300.5 | 231.3 | 116.1 | 7.3 | 53.4 | 170.1 | 116.8 | 113.3 | 10.4 |
| 4  | 313.4 | 503.4 | 190.0 | 416.4 | 24.6 | 102.7 | 148.2 | 45.5 | 128.1 | 5.5 | 124.3 | 192.8 | 68.5 | 158.5 | 12.2 |
| 5  | 492.2 | 620.9 | 128.7 | 581.4 | 15.9 | 57.1 | 310.4 | 253.4 | 154.8 | 92.9 | 32.8 | 124.0 | 122.8 | 76.5 | 17.1 |
| 6  | 545.0 | 583.6 | 38.6 | 569.4 | 7.1 | 97.6 | 291.8 | 194.2 | 282.8 | 18.8 | 70.3 | 111.0 | 40.7 | 86.2 | 7.9 |
| 7  | 170.7 | 559.5 | 388.8 | 450.9 | 25.0 | 49.9 | 264.0 | 214.1 | 125.7 | 7.8 | 73.1 | 176.9 | 103.8 | 143.7 | 7.0 |

| ID | Latent heat flux | ET-instantaneous (mm/hr) | ET-24 (mm/day) |
|----|------------------|--------------------------|----------------|
|    | Min | Max | Range | Mean | Std | Min | Max | Range | Mean | Std | Min | Max | Range | Mean | Std |
| 1  | 1.6 | 485.4 | 483.9 | 207.2 | 72.0 | 0.002 | 0.699 | 0.697 | 0.298 | 0.104 | 0.05 | 16.77 | 16.72 | 7.16 | 2.49 |
| 2  | 3.6 | 492.1 | 488.5 | 222.9 | 60.1 | 0.005 | 0.708 | 0.703 | 0.321 | 0.087 | 0.12 | 17.00 | 16.88 | 7.70 | 2.08 |
| 3  | 106.8 | 468.9 | 362.0 | 296.8 | 35.9 | 0.154 | 0.675 | 0.521 | 0.427 | 0.052 | 3.69 | 16.20 | 12.51 | 10.25 | 1.24 |
| 4  | 32.6 | 232.1 | 199.5 | 129.7 | 37.1 | 0.047 | 0.334 | 0.287 | 0.187 | 0.053 | 1.12 | 8.02 | 6.89 | 4.48 | 1.28 |
| 5  | 162.6 | 500.6 | 338.1 | 350.1 | 77.5 | 0.234 | 0.721 | 0.487 | 0.504 | 0.111 | 5.62 | 17.29 | 11.68 | 12.09 | 2.68 |
| 6  | 161.5 | 376.3 | 214.8 | 200.4 | 19.6 | 0.232 | 0.542 | 0.309 | 0.288 | 0.028 | 5.58 | 13.00 | 7.42 | 6.92 | 0.68 |
| 7  | 10.5 | 405.4 | 394.9 | 181.4 | 25.7 | 0.000 | 0.584 | 0.584 | 0.261 | 0.037 | 0.00 | 14.00 | 14.00 | 6.27 | 0.88 |
5. Conclusions

Estimated results of hourly and daily evapotranspiration based on SEBAL method and Landsat-8 imagery show a distinct pixel wise variation in the pattern of ET. The variation of ET for different land cover shows the hourly ET ranged from 0 at bare land and rock outcrops to 0.504 mm per hour at forest area, and the daily ET varied from 0.0 to 12.09 mm per day at the same land cover classes. The main land covers controlling the ET characteristics is the forests, which is covering about 51.6% of the study area. The least range of variation in the hourly ET values is appeared in the bare land area which is characterized by the absence of vegetation cover. The results show clear relation between land use/land cover and solar radiation parameters and impact of vegetation cover on the ET values in pixel wise domain.

Acknowledgements

Special thanks to University of Warsaw, Faculty of Geography and Regional Studies for their support in supplying research requirements. Thank you for the staff of Department of Geoinformatics Cartography and Remote Sensing for completion field validation for land use/land cover data.

References

[1] Bashir, M., Hata, T., Abdelhadi, A., Tanakamaru, H., and Tada, A. 2006. “Satellite-Based Evapotranspiration and Crop Coefficient for Irrigated Sorghum in the Gezira Scheme, Sudan.” Hydrology and Earth System Sciences Discussions 3: 793-817.
[2] Spiliotopoulos, M., Adaktylou, N., Loukas, A., Michalopoulou, H., Mylopoulos, N., and Toulis, L. 2013. “A Spatial Downscaling Procedure of MODIS Derived Actual Evapotranspiration Using Landsat Images at Central Greece.” In First International Conference on Remote Sensing and Geoinformation of Environment, International Society for Optics and Photonics, 879508-9.
[3] Trezza, R., Allen, R. G., and Tasumi, M. 2013. “Estimation of Actual Evapotranspiration along the Middle Rio Grande of New Mexico Using MODIS and Landsat Imagery with the METRIC Model.” Remote Sensing 5: 5397-423.
[4] Merlin, O., Chirouze, J., Olioso, A., Jarlan, L., Chehbouni, G., and Boulet, G. 2014. “An Image-Based Four-Source Surface Energy Balance Model to Estimate Crop Evapotranspiration from Solar Reflectance/Thermal Emission Data (SEB-4S).” Agricultural and Forest Meteorology 184: 188-203.
[5] Hailegiorgis, W. S. 2006. “Remote Sensing Analysis of Summer Time Evapotranspiration Using SEBS Algorithm.” ITC, Enschede.
[6] Hong, S.-H., Hendricks, J. M. H., and Borchers, B. 2009. “Up-scaling of SEBAL Derived Evapotranspiration Maps from Landsat (30 m) to MODIS (250 m) Scale.” Journal of Hydrology 370: 122-38.
[7] Venturini, V., Krepper, C., and Rodriguez, L. 2012. “Evapotranspiration Estimation Based on the Complementary Relationships.” In Evapotranspiration—Remote Sensing and Modeling, edited by Irmak, A. Intechweb.org: Rijeka, Croatia, 19-40.
[8] Courault, D., Seguin, B., and Olioso, A. 2005. “Review on
Estimation of Evapotranspiration from Remote Sensing Data: From Empirical to Numerical Modeling Approaches.” Irrigation and Drainage Systems 19: 223-49.

Bastiaanssen, W., Pelgrum, H., Wang, J., Ma, Y., Moreno, J., Roerink, G., and Van der Wal, T. 1998. “A Remote Sensing Surface Energy Balance Algorithm for Land (SEBAL): Part 2: Validation.” Journal of Hydrology 212: 213-29.

Tasumi, M., Trezza, R., Allen, R. G., and Wright, J. L. 2003. “US Validation Tests on the SEBAL Model for Evapotranspiration via Satellite.” In Proceedings of 54th IEC Meeting of the International Commission on Irrigation and Drainage (ICID) Workshop Remote Sensing of ET for Large Regions.

Bastiaanssen, W., Noordman, E., Pelgrum, H., Davids, G., Thoreson, B., and Allen, R. 2005. “SEBAL Model with Remotely Sensed Data to Improve Water-Resources Management under Actual Field Conditions.” Journal of Irrigation and Drainage Engineering 131: 85-93.

Ruhoff, A. L., Paz, A. R., Collischonn, W., Aragao, L. E., Rocha, H. R., and Malhi, Y. S. 2012. “A MODIS-Based Energy Balance to Estimate Evapotranspiration for Clear-Sky Days in Brazilian Tropical Savannahs.” Remote Sensing 4: 703-25.

Mirek, Z., and Glowaciński, Z. 1996. “Przyroda Tatrzanskiego Parku Narodowego.” wspólpr. Polska Akademia Nauk. Instytut Botaniki im. W. Szafera.

Guzik, M., and Skawiński, P. 2009. “Applying Geomatics to Determination of Landscape Altitudinal Zones in the Mountains.” Landform Analysis 11: 25-32.

Birkenmajer, K. 2012. “Geology of the Lower Subalpine Nappe, Kopy Sołtyśie Area, Eastern Tatra Mts (West Carpathians, Poland).” Studia Geologica Polonica 135: 55-116.

Ksiazkiewicz, M. 1956. “Geology of the Northern Carpathians.” Geologische Rundschau 45: 369-411.

Mirek, Z., and Piekons-Mirkowa, H. 1992. “Flora and Vegetation of the Polish Tatra Mountains.” Mountain Research and Development: 147-73.

Mather, P., and Tso, B. 2009. Classification Methods for Remotely Sensed Data. CRC press.

ERSOS, LANDSAT 8- Path: 188 Row: 26 for Scene: LC81880262013251LGN00, Geospatial Data Presentation Form: Remote-Sensing Image, in, Sioux Falls, South Dakota, USA 2013.

Chander, G., and Markham, B. 2003. “Revised Landsat-5 TM Radiometric Calibration Procedures and Postcalibration Dynamic Ranges.” Geoscience and Remote Sensing. IEEE Transactions 41: 2674-7.