A distributed thermodynamic model for energy and mass balance computation: FEST–EWB

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Abstract:

This article presents the development of distributed thermodynamic model for energy and mass balance computation between soil surface and shallow atmospheric layers and its inclusion into the hydrological model FEST–EWB (Flash–flood Event–based Spatially distributed rainfall–runoff Transformation–Energy Water Balance). This model is also thought for a synergic use of hydrological model with remote sensing data. In particular, the energy budget is solved looking for the representative thermodynamic equilibrium temperature (RET) defined as the land surface temperature (LST) that closes the energy balance equation for any pixel of basin surface. So using this approach, through the system between the mass and energy equations, soil moisture (SM) is linked to the latent heat flux (LE) and then to LST. The RET thermodynamic approach solves most of the problems of the actual evapotranspiration (ET) and SM computation. In fact, it permits to avoid computing the effective ET as an empirical fraction of the potential one. This approach, based on the RET, has been tested at field scale (10 ha) with energy fluxes and LST measured with an eddy covariance station in Landriano (Italy). Copyright © 2010 John Wiley & Sons, Ltd.

KEY WORDS energy water balance model; representative thermodynamic equilibrium temperature; land surface temperature; energy fluxes

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INTRODUCTION

Hydrological models can give an important contribution to quantify mass and energy fluxes at the basin and irrigation district scale (Famiglietti and Wood, 1994; Montaldo and Albertson, 2001; Gurtz *et al.*, 2002; Ravazzani *et al.*, 2008; Rabuffetti *et al.*, 2008).

Nevertheless, their application, both in operative and scientific research, is limited by the difficulties to verify evaporative fluxes and soil water content at the basin scale. In fact, calibration and validation of distributed models generally depend on comparison between simulated and observed discharges at the available rivers cross-sections.

Soil moisture (SM), which is recognized as the key variable in these hydrologic energy water balance models, respect to this role is most of the time confined to an internal numerical model variable (Dooge, 1986) and it seems that the link between internal variables of the processes [e.g. SM, land surface temperature (LST) and evapotranspiration (ET) fluxes] and external variables, which are the outputs of the model (e.g. discharge measurements), is not yet resolved. Consequently, the physically based distributed models risk to be still considered as ‘Black-Box’ models.

These problems drove the scientific community to the use of hydrologic modelling in conjunction with remote sensing data. In particular, there was a development of

distributed hydrological models with mass and energy balance equations (‘Land surface model’, LSM) for water content and ET estimations which use remote sensing data of LST, a connected variable to SM, as input parameters, validation terms or assimilation variables (Noihlan and Planton, 1989; Famiglietti and Wood, 1994; Anderson *et al.*, 1997; Bastiaanssen *et al.*, 1998; Montaldo and Albertson, 2001; Su, 2002; Corbari *et al.*, 2008, 2010).

This approach based on LST seems to solve many limitations and difficulties of the previous technology based on micro-wave satellite images (Giacomelli *et al.*, 1995; Mancini *et al.*, 1999). In fact, promising results are now coming using both hydrological modelling and thermal infrared satellite images available from sensors like MODIS, ASTER and SEVIRI on board the main operative satellites (Anderson *et al.*, 1997; Bastiaanssen *et al.*, 1998; Minacapilli *et al.*, 2009).

In particular, as stated by Su (2002) there exist three different types of energy balance models for ET estimates that use satellite and ground data as inputs: (1) residual approach, (2) to compute all the terms of the energy balance and (3) semi-empirical approach.

However, the models with a residual flux in the energy balance are more common in the scientific community and they can be subdivided into: (1) ‘one-source’ models that do not distinguish between evaporation and transpiration, such as the SEBAL model (Bastiaanssen *et al.*, 1998) and the SEBS model (Su, 2002), that computes the latent heat flux (LE) as a residual term of the energy

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budget and (2) 'two-source' models that instead consider evaporation and transpiration separately, such as the TSEB model (Norman *et al.*, 1995) where the LE is the residual or the FORCE–RESTORE model (Montaldo and Albertson, 2001) that computes the ground heat flux as a residual flux.

As stated by Kalma *et al.* (2008), the methods that compute all the terms of the energy balance are widely diffuse, but most of them are coupled with data assimilation techniques (e.g. LST from satellite data) to update the state variables. Instead, low efforts have been made to use LST for the calibration of some parameters of the models. Moreover, these models (e.g. VIC, Liang *et al.*, 1994; or TOPLATS, Famiglietti and Wood, 1994) do not solve simultaneously the energy and mass balance equations.

In this context, this article has as objective the development of a thermodynamic model for energy and mass balance computation and its inclusion into the distributed hydrological model FEST–EWB (Flash-flood Event-based Spatially distributed rainfall–runoff Transformation–Energy Water Balance) (Corbari, 2010). In particular, the approach of solving the energy budget between soil surface and shallow atmospheric layers looking for the defined LST that closes the energy balance equation in any pixel of the basin surface (hereafter called representative equilibrium temperature, RET) is developed for local scale as for large-scale use.

Particularly in this article, the validation of this new version of the FEST–EWB model is performed at local scale with ground measurements of LST and energy fluxes.

Future developments of this work will be the validation of the RET approach at basin scale by comparing in each pixel of the basin of the simulated RET (internal variable) with LST from satellite images together with the traditional calibration based on the comparison between simulated and observed discharges (external variable) at the control cross-sections of the rivers.

THE DISTRIBUTED HYDROLOGICAL MODEL: FEST–EWB SCHEME

FEST–EWB is a distributed hydrological energy water balance model (Corbari *et al.*, 2008; Corbari, 2010) and it is developed starting from the FEST–EWB (Mancini, 1990; Rabuffetti *et al.*, 2008; Ravazzani *et al.*, 2008). FEST–EWB computes the main processes of the hydrological cycle: ET, infiltration, surface runoff, flow routing, subsurface flow and snow dynamics (Corbari *et al.*, 2009). In the FEST–EWB, the energy balance module is introduced. The model is distributed so that the computation domain is discretized with a mesh of regular square cells in which every parameter is defined or calculated.

The variables of the model are listed in Table I including symbols, meanings and units of measurement.

Table I. List of symbols, their meaning and unit

Symbol	Meaning	Unit
A	Radiation reflected from neighbouring terrain	—
B	Solar azimuth	—
C	Angle between the sunbeam direction and the perpendicular to the ground	—
c_b	Specific heat capacity of biomass	$\text{J kg}^{-1} \text{K}^{-1}$
c_p	Specific heat of humid air	$\text{MJ kg}^{-1} \text{K}^{-1}$
c_s	Specific heat capacity of soil	$\text{J kg}^{-1} \text{K}^{-1}$
c_w	Specific heat capacity of water	$\text{J kg}^{-1} \text{K}^{-1}$
d	Zero plane displacement height	m
DF	Scattered radiation	—
E	Mountain aspect	—
e^*	Saturation vapour pressure	Pa
e_a	Vapour pressure	Pa
ET	Evapotranspiration rate	mm
FC	Field capacity	—
f_v	Vegetation fraction	—
F_{CO_2}	Photosynthesis flux	W m^{-2}
G	Soil heat flux	W m^{-2}
g^{term}	Soil thermal conductivity	$\text{W m}^{-1} \text{K}^{-1}$
h	Solar elevation angle	—
$H_s - H_c$	Sensible heat fluxes for bare soil (s) and for canopy (c)	W m^{-2}
H_w	Sensible heat flux over water	W m^{-2}
I_c	Incoming solar radiation reaching the ground in the normal direction	—
I_0	Extraterrestrial radiation at the top of the atmosphere	—
k	Von Karman constant	—
K_i	Fraction of scattered radiation	—
L	Obukhov length	m
LAI	Leaf area index	—
$LE_s - LE_c$	Latent heat fluxes for bare soil (s) and for canopy (c)	W m^{-2}
LST	Land surface temperature	K
LE_w	Latent heat flux over water	W m^{-2}
m_b	Mass of biomass	kg m^{-2}
m_w	Mass of water	kg m^{-2}
p	Reduction fraction for sky covered with clouds	—
P	Precipitation rate	mm
PE	Drainage flux	mm
Q	Direct radiation component	—
Q_a	Actual direct radiation	—
r	Albedo	—
R	Runoff flux	mm
R^*	Theoretical radiation observed at the ground level	—
r_a	Aerodynamic resistance for vegetation	s m^{-1}
r_{abs}	Aerodynamic resistance for bare soil	s m^{-1}
r_c	Canopy resistance	s m^{-1}
RET	Representative equilibrium temperature	K
RH	Air humidity	—

In Figure 1, the scheme of the complete hydrological model is reported. In particular, the input parameters of the model, surrounded by a rectangle with dashed lines, are: (1) the atmospheric forcings, such as air temperature,

Table I. (Continued)

Symbol	Meaning	Unit
Rn	Net radiation	$W m^{-2}$
R_s	Incoming short-wave radiation	$W m^{-2}$
r_s	Soil resistance	$s m^{-1}$
$r_{s \min}$	Minimum stomatal resistance	$s m^{-1}$
s	Atmosphere optical depth	—
s_0	Optical depth at sea level (Kreider and Kreith, 1975)	—
S_{air}	Air enthalpy changes	$W m^{-2}$
S_{canopy}	Crop enthalpy changes	$W m^{-2}$
S_{soil}	Soil surface layer heat flux	$W m^{-2}$
SM	Soil moisture	—
S_w	Storage term of water	$W m^{-2}$
T_a	Air temperature	K
T_{soil}	Temperature below the first layer of soil dz	K
u	Horizontal wind velocity	$m s^{-1}$
WP	Wilting point	—
WT_{prof}	Water temperature at a fixed depth	K
z	Height above sea level	M
z_{oh}	Roughness length governing transfer of heat and vapour	M
z_h	Height of humidity measurements	M
z_{om}	Roughness length governing momentum transfer	M
z_m	Height of wind measurements	M
α	Mountain slope	—
γ	Psychrometric constant	$Pa^{\circ}C^{-1}$
η	Nash and Sutcliffe index	—
λ	Latent heat of vapourization	$MJ kg^{-1}$
ξ_c	Atmosphere emissivity	—
ξ_s	Soil emissivity	—
ρ_a	Air density	$kg m^{-3}$
ρ_s	Soil density	$kg m^{-3}$
ρ_w	Water density	$kg m^{-3}$
σ	Stefan–Boltzmann constant	$W m^{-2} K^{-4}$
ΔRET	Difference of RET in Δt	—
Δt	Time step	S
ΔT_a	Difference of air temperature in Δt	$^{\circ}C$
ΔT_{soil}	Difference of soil temperature in Δt	—
$\Delta W/\Delta t$	Energy storage terms	$W m^{-2}$
Δz_{air}	Height of the measurement instrument of T_a	m
Δz_{soil}	Soil heat flux plate depth	m
Δz_w	Water depth	m
ψ	Angle between the point with maximum elevation in the direction of the solar beam	—
ψ_h	Functions for atmospheric stability governing transfer of heat and vapour	—
ψ_m	Functions for atmospheric stability governing momentum transfer	—

incoming short-wave radiation, wind velocity, precipitation, air humidity from ground stations, that need to be interpolated or from meteorological models, (2) the digital elevation model (DEM), (3) the soil parameters in distributed maps, such as the saturated hydraulic conductivity, the field capacity or the soil depth and (4) the vegetation parameters, such as leaf area index and vegetation height, which can be retrieved from ground or satellite

Figure 1. FEST-EWB scheme

Figure 2. Energy and mass fluxes calculated in FEST-EWB

data. All the processes computed by the model are surrounded by a rectangle with continuous lines, while the outputs of the model, surrounded in Figure 1 by a circle, can be distributed values like LST and ET or/and local values, such as the discharge in the control cross-section of the river.

The core of the model is the system between the water and energy balance equations that are linked through ET. The terms of these equations are reported in Figure 2.

In particular, in the energy balance between soil surface, vegetation and low atmosphere, all the terms are computed and the equation is solved looking for the RET that is the LST that closes the energy balance equation.

So using this approach, SM is linked to the LE and then to RET. In fact, LE is computed as a function of the LST that closes the energy budget, but in addition the LE, as effective ET, is also a term of the mass balance equation which computes at every time step the

evolution of SM. In fact, more SM is available in the soil, more ET is present, but also less energy is available for heating the surface. So through this system of energy and mass balance equations, LST variability can influence SM changes significantly.

Moreover, the RET thermodynamic approach solves most of the problems of the actual ET and SM computation. In fact, it permits the model to avoid computing the effective ET as an empirical function of potential ET, such as the beta function developed by Parlange *et al.* (1999).

The water balance

SM evolution for the generic cell at positions i and j is described by the water balance equation:

$$\frac{\partial \text{SM}_{i,j}}{\partial t} = \frac{1}{dz_{i,j}} (P_{i,j} - R_{i,j} - \text{PE}_{i,j} - \text{ET}_{i,j}) \quad (1)$$

The water balance equation is computed in each cell of the domain characterized with a single layer of soil depth dz .

SM in cells covered by snow is assumed not varying with time.

The energy balance

Land surface energy balance is a reformulation of the first law of thermodynamics, which deals with the energy conservation where the total energy of a system remains the same even if energy transformations occur.

The complete energy balance equation at the ground surface for the generic cell at positions i and j is expressed as

$$\begin{aligned} \text{Rn}_{i,j} - G_{i,j} - (H_s + H_c)_{i,j} - (\text{LE}_s + \text{LE}_c)_{i,j} \\ = \frac{\Delta W}{\Delta t_{i,j}} \end{aligned} \quad (2)$$

These $\Delta W/\Delta t$ energy storage terms are often negligible, especially at basin scale with a low spatial resolution; instead at local scale the contribution of these terms could be significant (Meyers and Hollinger, 2004; Jacobs *et al.*, 2008; Corbari, 2010).

The terms of the energy balance equation are described in the following section.

Net radiation. The net radiation is the algebraic sum of the incoming and outgoing short- and long-wave radiation:

$$\text{Rn} = R_s(1 - r) + \xi_c \sigma (T_a)^4 - \xi_s \sigma (\text{RET})^4 \quad (3)$$

In literature, several equations exist for the description of the atmosphere emissivity, ξ_c , considering clear and cloudy skies with different cloud cover fraction (Sedlar and Hock, 2009).

Short-wave net radiation, $R_s(1 - r)$, involved in the evaporative processes is calculated considering the effect of topography (Mancini *et al.*, 2005). The radiation balance at basin surface varies with latitude, season, time

Figure 3. Radiation affected by topography

and presence of clouds; moreover, in mountain basins, topographic characteristics, such as slope and aspect, may have a relevant role. In fact, the topography affects the radiation field in three ways (Figure 3): (1) modulating the actual energy flux according to the relative position of the ground surface with respect to the sun, (2) reducing radiation because of shadowing effect of the higher ridges of mountains and (3) increasing net radiation by the fraction reflected from neighbouring terrains.

The incident short-wave radiation can be expressed as

$$R_s = Q_a + \text{DF} + A \quad (4)$$

In clear sky conditions, the incoming solar radiation reaching the ground in the normal direction is

$$I_c = I_0 \exp \left[\frac{-s}{\sin(h)} \right] \sin(h) \quad (5)$$

where the atmosphere optical depth can be computed as $s = s_0 P_z/P_0$ and P_z/P_0 is a correction factor that takes into account the difference in atmospheric pressure between sea level (P_0) and actual elevation (P_z):

$$\frac{P_z}{P_0} = \left(\frac{288 - 0.0065 \cdot z}{288} \right)^{5.256} \quad (6)$$

The scattered radiation for clear sky condition, D , is

$$D = k_b (I_0 \sin(h) - I_c) \quad (7)$$

where k_b varies from 0.2 to 0.6 according to the sky brightness.

In clear sky condition, the theoretical radiation observed at the ground level is expressed by

$$R^* = I_c + D \quad (8)$$

The presence of clouds or natural obstacles reduces the direct radiation I_c and modifies the scattered one so that R^* can be reduced to a minimum fraction, p , of R^* (usually $p = 0.22$ is referred to a transmissivity coefficient for 8/8 stratocumulus cloud cover; Male and Granger, 1981). When the observed radiation is less than or equal to $p \times R^*$, then the radiation is considered totally scattered (sky totally covered with clouds). Otherwise, the fraction of scattered radiation, K_t , is computed as

$$K_t = \frac{R^* - R_s}{[(1 - p)R^*]} \quad (9)$$

Finally, the component of the actual scattered radiation, DF, can be computed as

$$DF = \min[R_s \cdot D \cdot (1 - K_t) + R_s \cdot K_t] \quad (10)$$

and the direct radiation component, Q , as

$$Q = R_s - DF \quad (11)$$

Direct radiation component, Q , is affected by topographic characteristics, such as slope, α , and aspect, E . Actual direct radiation, Q_a , is related to the sun elevation as

$$Q_a = \frac{Q \cos(C)}{\sin(h)} \quad (12)$$

The angle (Figure 3) between the sunbeam direction and the perpendicular to the ground is evaluated as

$$\cos(C) = \cos(h) \cdot \sin(\alpha) \cdot \cos(B - E) + \sin(h) \cdot \cos(\alpha) \quad (13)$$

The radiation reflected from the neighbouring terrain is

$$A = Q_a \cdot r \cdot (1 - f_\alpha) \quad (14)$$

where

$$f_\alpha = 1 - \frac{\alpha}{180^\circ}.$$

A module taking into account the shadow effect induced by topography was developed. The algorithm calculates the angle, ψ , between the point with maximum elevation in the direction of the solar beam, denoted by coordinates x_m , y_m and z_m , and the examined cell, denoted by coordinated x_0 , y_0 and z_0 (Figure 3):

$$\psi = \arctan \left[\frac{(z_m - z_0)}{\sqrt{(x_m - x_0)^2 + (y_m - y_0)^2}} \right] \quad (15)$$

If ψ is higher than the sun elevation, the cell is shadowed and the incident short-wave radiation is constituted by scattered component only:

$$R_s = DF \quad (16)$$

Soil heat flux. The soil heat flux is the heat changed for conduction with the subsurface soil and it is evaluated as

$$G = \left(\frac{g_{\text{term}}}{dz} \right) (RET - T_{\text{soil}}) \quad (17)$$

The soil thermal conductivity, which depends on the soil water tension, is evaluated with the McCumber–Pielke equation that computes the soil water tension as a function of SM (McCumber and Pielke, 1981; Peters-Lidard *et al.*, 1998).

Sensible heat flux. The sensible heat flux, which is heat energy transferred between the surface and air when their temperatures are different, is considered for bare soil, H_s , and for vegetation, H_c . The cell of the computational domain is characterized by a vegetation fraction, f_v , to

discriminate the percentage of vegetation coverage. The equation of this vertical flux is

$$H_s + H_c = (1 - f_v) \frac{\rho_a c_p}{r_{\text{abs}}} (RET - T_a) + f_v \frac{\rho_a c_p}{r_a} (RET - T_a) \quad (18)$$

The aerodynamic resistance regulates the transfer of heat and water vapour from the evapotranspiring surface into the air above the canopy. Correction functions for atmospheric stability or instability are included using the Thom model (Thom, 1975).

Aerodynamic resistance for vegetation fraction can be calculated using:

$$r_a = \frac{\left[\ln \left(\frac{z_m - d}{z_{om}} \right) - \Psi_m \left(\frac{z_m - d}{L} \right) \right] \left[\ln \left(\frac{z_h - d}{z_{oh}} \right) - \Psi_h \left(\frac{z_h - d}{L} \right) \right]}{k^2 u} \quad (19)$$

where z_m , d , z_{om} , z_h and z_{oh} are functions of the vegetation height. In case of bare soil condition, the aerodynamic resistance, r_{abs} , should be calculated using the same Equation (19), but z_m , d , z_{om} , z_h and z_{oh} are calculated using vegetation height as the zero value.

Latent heat flux. The LE, which is the heat used in the phase change from a liquid to a gas, is considered for bare soil, LE_s , and for canopy, LE_c , and in the FEST-EWB is computed as

$$LE_c + LE_s = f_v \left(\frac{\rho_a c_p}{\gamma(r_a + r_c)} \right) (e^* - e_a) + (1 - f_v) \frac{\rho_a c_p}{\gamma(r_{\text{abs}} + r_s)} (e^* - e_a) \quad (20)$$

The saturation vapour pressure, e^* , is computed as a function of the RET while the vapour pressure, e_a , as a function of T_a .

The canopy resistance, which describes the resistance of vapour flow through the transpiring crop, is expressed as (Jarvis, 1976)

$$r_c = \frac{r_{\text{smin}} (FC - WP)}{LAI (\theta - WP)} \quad (21)$$

The soil resistance is the resistance at the evaporating soil surface (Sun, 1982) and is computed as

$$r_s = 3.5 \left(\frac{\theta_{\text{sat}}}{\theta} \right)^{2.3} + 33.5 \quad (22)$$

The latent heat of vapourization and the water density link the LE with the ET (ms^{-1}) as

$$LE = \lambda \rho_w ET \quad (23)$$

Storage fluxes. The storage fluxes, which in literature have been demonstrated to be significant, can be calculated only when local scale simulations are performed where a micrometeorological station is available so that particular measurements are obtainable, such as the CO_2 flux or values of air temperature at different height.

Figure 4. Shape of the complete energy balance equation solved with the Newton–Raphson method and the zero of the function

So if a local scale analysis is performed, the photosynthesis flux (F_{CO_2}), the crop and air enthalpy changes (S_{canopy} and S_{air}) and the soil surface layer heat flux (S_{soil}) are the additional fluxes (Wm^{-2}) considered in the energy balance:

$$\frac{\Delta W}{\Delta t_{i,j}} = F_{CO_{2i,j}} + S_{canopyi,j} + S_{airi,j} + S_{soili,j} \quad (24)$$

The photosynthesis flux, that is the change in the Gibbs free energy, is calculated with the conversion from the measured flux of $1 \text{ mg CO}_2 \text{ m}^{-2} \text{ s}^{-1}$ to 11 W m^{-2} (Nobel, 1974).

The air enthalpy is evaluated as

$$S_{air} = \frac{\Delta T_a \rho_a c_p}{\Delta t} \Delta z_{air} \quad (25)$$

The canopy enthalpy is computed only over a fixed vegetation height:

$$S_{canopy} = \frac{\Delta RET(m_w c_w + m_b c_b)}{\Delta t} \quad (26)$$

A similar approach is used for the heat flux in the soil surface layer:

$$S_{soil} = \frac{\Delta T_{soil}(SMm_w c_w + \rho_s c_s)}{\Delta t} \Delta z_{soil} \quad (27)$$

This soil energy flux is the same term that is added in the post-processing of the analysed data to the soil heat flux measured by a heat flux plate.

Solution scheme. All the terms of the energy budget, as seen in the precedent sections, depend on RET and so the energy balance equation can be solved

looking for the RET that closes the equation with the well known Newton–Raphson method. The iterative procedure implemented in the FEST–EWB model is

$$LST_n = LST_{n-1} + \frac{f_t(LST_{n-1})}{f'_t(LST_{n-1})} \quad (28)$$

where LST_n is the actual value, LST_{n-1} the value at the previous iteration, $f_t(LST_{n-1})$ the energy balance function and $f'_t(LST_{n-1})$ its derivative. The solution is acceptable when $\left| \frac{f_t(LST)}{f'_t(LST)} \right| < \text{tolerance}$ and $f_t(LST) < \text{tolerance}$, with tolerance equal to 0.001.

The Newton–Raphson method is the best known method for finding successively better approximations to the zeros of a function and it converges often quickly, especially if the iteration begins sufficiently near the solution. For the correct interpretation of the mathematical results of the method, the shape of the complete energy balance equation has been analysed. Different possible atmospheric and soil coverage conditions have been considered and the function always has only one zero. In the following text, the energy balance equation *versus* LST is reported for daily values in a bare soil condition and in a vegetative period and for night values in a bare soil condition (Figure 4).

Energy balance equation for water surface. For water bodies, such as lake or paddy, the energy water balance equation is introduced and is solved, as for the soil surface energy budget, with the Newton–Raphson method:

$$Rn - H_w - LE_w = S_w \quad (29)$$

Figure 5. Meteorological variables used as input into the model

The LE is considered in its potential form, while G (W m^{-2}) is often negligible. The storage term of water is computed as

$$S_w = \frac{(LST - WT_{\text{prof}})\rho_w c_w}{\Delta t} \Delta z_w \quad (30)$$

SYNTHETIC TEST CASES

Synthetic test cases are presented to show the functioning of the RET approach at local scale. A data set of meteorological conditions is created by repeating single day's measurements of incoming short-wave radiation (R_s), air humidity (RH), wind velocity (u) and air temperature (T_a). This data set is reported in Figure 5. During the simulations the rain (P) is set to zero.

Four different cases for bare soil and vegetation with different SM conditions are considered: (1) bare soil with SM set as initial condition equal to 0.4, (2) bare soil with SM equal to 0.18 (3) vegetation of 2-m height with SM equal to 0.4 and (4) vegetation of 2-m height with SM equal to 0.18.

In Figure 6, the energy fluxes (e.g. net radiation, latent and sensible heat fluxes and soil ground heat flux), LST and SM are reported for each test case. In particular, the difference between bare soil and vegetation is correctly visible. In fact for the two bare soil cases 1 and 2, the sensible heat flux is predominant on the LE, while it is the opposite in both the vegetation cases 3 and 4. Moreover, if the LST is considered, in both the bare soil cases the daily peak of LST is higher than in both the vegetation cases of about 10°C .

Instead looking only at vegetation conditions, case 4 has lower soil water content than case 3 which corresponds to a lower LE, a higher LST and consequentially a higher sensible heat flux. These results can also be found from the comparison between bare soil case 2 with lower SM initial state and case 1 with higher SM.

MODEL VALIDATION AGAINST OBSERVED DATA

FEST-EWB model based on the RET approach can now be validated using LST measured by ground radiometer in any pixel of the analysis domain.

In this article, the model is validated at the study site of Landriano (Italy), an experimental field of maize, where an eddy covariance station has been positioned for monitoring energy and mass exchanges. In particular, the considered data set is composed of LST and net radiation measured from the CNR1 by Kipp and Zonen radiometer, latent and sensible heat fluxes are computed from the covariance between the vertical wind velocity and the relative air humidity and the air temperature measured from the sonic anemometer (Young 81000 by Young) and the gas analyser (LICOR 7500 by LICOR), while the ground heat flux is measured by two thermocouples ELSI and a heat flux plate (HFP01 by Hukseflux). These data are recorded at 30-min time step from 13 March 2006 to 11 October 2006, when the maize was harvested. Maize was planted on 1 June 2006.

The validation process is based on the comparison between the simulated energy fluxes and LST and the measured ones by the micrometeorological station. More details about Landriano eddy covariance stations, its entire instrumentation, its data set and the analysis on the quality of the measured data can be found in Corbari (2010).

Energy fluxes

The measured fluxes are compared with the simulated ones and a good accuracy is reached for the net radiation and the latent and sensible heat fluxes. Instead, the soil heat flux is reproduced in disagreement with the observed data due to representativeness problem of the measured G (Kustas *et al.*, 2000) (Figure 7). These results are also confirmed from a statistical analysis (Table II) looking for the minimization of the root mean square error (RMSE) and the maximization of the efficiency of the Nash and Sutcliffe index, η (Nash and Sutcliffe, 1970).

In Table II, the angular coefficients of the linear regression curves between simulated and observed energy fluxes and their R^2 are also reported. Also, the cumulated fluxes are well reproduced.

LST from ground data

LSTs measured from the radiometer at the eddy covariance station are compared with RETs simulated from FEST-EWB and a mean difference over the entire

Figure 6. Test cases for bare soil and vegetation with different SM conditions

period of simulation of -0.7°C is reached. In Table II, some statistical results are reported, confirming the good capability of the model in reproducing observed LSTs. In Figure 8, the scatter plots between the simulated RET and the observed LST are reported showing similarity between the two variable values.

CONCLUSIONS

This article presented an approach for energy water balance computation based on the thermodynamic equilibrium implemented in a distributed hydrological model FEST-EWB for a synergic use of hydrological model

with remote sensing and ground data. The energy budget is solved looking for the representative thermodynamic equilibrium temperature, RET, defined as the LST that closes the energy balance equation for any pixel of basin surface.

The energy approach method based on the RET temperature seems to provide consistent results for vegetated and non-vegetated areas. In fact, the synthetic test cases showed the correct functioning of the model for different conditions. Moreover, the comparison between observed LST from ground radiometer at Landriano station and simulated RET confirms the goodness of the approach and a good capability in representing the observed data

Figure 7. Comparison between the simulated and measured energy fluxes

Table II. RMSE, the Nash and Sutcliffe index, the angular coefficient of the linear regression and the errors on the cumulated total energy for the energy fluxes and LST

	η	RMSE	Angular coefficient	R^2	Total energy (%)
Rn	0.96	38.3	$y = 0.95 x$	0.97	2.4
LE	0.75	54.8	$y = 0.87 x$	0.70	3.4
H	0.71	29.6	$y = 0.80 x$	0.66	-1.3
G	0.68	22.5	$y = 0.84 x$	0.80	7.2
LST	0.87	3.2 (°C)	$y = 0.98 x$	0.89	—

is also shown with the comparison between the modelled and observed energy fluxes at the eddy covariance station.

The possibility given by this approach to compare observed LST in a quantitative way from ground radiometers and computed from model offers the possibility to test this model in the future over larger area

Figure 8. Comparison between the simulated and the observed LST

validating it with LST from satellite images. Moreover, this way of modelling the soil water balance seems to solve the problems of controlling the hydrological models only with external variable, such as discharge. In fact,

the computation of LST at pixel scale can be considered as an increase of points where fluxes can be controlled in addition to the traditional gauges stations at the basin outlet.

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