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ABSTRACT

Single-source energy balance models are simple and particularly suited to assimilate mixed pixel remote sensing data. Mixed pixels are made up of a combination of two main elements, the soil and the vegetation. The use of single-source models implies that the reference temperature for the estimation of convective fluxes, the aerodynamic temperature, is linked to the available remotely sensed surface temperature. There are many relationships relating both temperatures in the literature, but few that try to find objective constraints on this link. These relationships account for the difference between both temperatures by dividing the roughness length for thermal turbulent transport by an expression known as "radiometric kB^{-1}", which depends mostly on Leaf Area Index (LAI). Acknowledging that the two temperatures should be similar for bare soil and high LAI conditions, we propose an empirical relationship between LAI and the ratio of the difference between the aerodynamic and the air temperatures and the difference between the surface and the air temperatures, also known as "β function". Nine datasets obtained in agricultural areas (four in south western France near Toulouse, four in south eastern France near Avignon, one in Morocco near

26 Marrakech) are used to evaluate this new relationship. They all span the entire cropping

27 season, and LAI values range from 0 to about 5. This new expression of the β function is then

28 compared to the β function retrieved from measured sensible heat flux and in-situ radiometric

29 measurements as well as the β function simulated by a two-source SVAT model (ICARE). Its

30 performance in estimating the sensible heat compares well to other empirical or semi-

31 empirical functions, either based on a β function or a radiometric kB^{-1}.

32

33 1. INTRODUCTION

34 Assessing the turbulent fluxes of latent and sensible heat at the land surface is a crucial issue

35 for both water resource management (computation of evapotranspiration) and meteorological

36 forecasting (evolution of the planetary boundary layer). In order to compute these fluxes at a

37 suitable spatial scale, estimation methods based on the use of remote sensing data are

38 favoured. There is a large array of evapotranspiration estimation methods that use as input or

39 constraint remotely sensed variables such as the NDVI (Courault et al., 2005; Gowda et al.,

40 2008; Olioso et al., 2005). Evapotranspiration in potential conditions (i.e. not water limited)

41 can be assessed with relatively good precision from in-situ (Cleugh et al., 2007) or remote

42 sensing derived (Venturini et al., 2008) meteorological data, and, very often, NDVI. But when

43 water stress occurs the latent and sensible heat fluxes are more difficult to assess. In those

44 cases, there is a tight coupling between the evapotranspiration and the radiative surface

45 temperature and, consequently, methods based on remote sensing data in the Thermal Infra

46 Red (TIR) domain are favoured (Kalma et al., 2008). Those methods often compute the

47 instantaneous latent heat flux as the residual of the energy budget and, in most cases, an

48 expression of each individual term of the energy budget is proposed (Boulet et al., 2007):

49 $$Rn - G = H + LE \tag{1}$$

50   While net radiation Rn and, to a lesser extent, soil heat flux G can be expressed directly as a

51   function of the radiative surface temperature, the turbulent fluxes H (sensible heat) and LE

52   (latent heat) depend on a mixed-surface (soil, air and vegetation) temperature source T_{aero}:

53    $$H = \rho c_p \frac{T_{aero} - T_{air}}{r_{a,h}}$$

(2)

54   where ρ is the air density, c_p the specific heat of air at constant pressure, T_{air} is the air

55   temperature at a reference level above the canopy and $r_{a,h}$ the aerodynamic resistance for heat

56   exchange. The temperature T_{aero} called aerodynamic temperature represents the average

57   temperature of the air in the vicinity of the vegetation elements within the canopy, at the

58   height of the aerodynamic level (defined as the sum of the displacement height and the

59   roughness length for momentum). There is no measuring device for this temperature, which is

60   usually inverted from turbulent flux measurements. Moreover, it can be significantly different

61   from the ensemble directional radiometric surface temperature T_{surf} (as defined by Norman

62   and Becker, 1995) which is usually derived from brightness temperature measurements made

63   by a thermo-radiometer or infrared thermometer at nadir or at a specific view angle (Kustas

64   and Anderson, 2009; Kustas et al., 2007; Stewart et al., 1994) and which is used to assess

65   energy balance from remote sensing data. In general, the relationship between aerodynamic

66   and surface temperatures is obtained through the use of a dual-source energy balance when

67   either the vegetation or the soil bulk skin temperatures are known (Lhomme et al., 2000).

68   Retrieving surface temperature for each of the different components of the surface (mostly

69   soil and plants) is difficult, especially with current remote sensing platforms (Jia et al., 2003).

70   The use of single-source models is therefore favoured over dual-source models to estimate

71   pixel average turbulent fluxes from a mixed-pixel radiometric temperature. For those models,

72   there is a need to develop robust yet simple methods to relate the aerodynamic temperature to

73   the surface temperature. This has been subject of debate for a long time (e.g. see Carlson et

74   al., 1995; Kalma and Jupp, 1990). Many formulations exist in the literature, and a

75   comprehensive terminology and conversion formulae are proposed by Matsushima (2005).

76   Historically, most expressions governing the relationship between the aerodynamic and the

77   surface temperatures have been built on an analogy between wind and temperature profiles

78   within the canopy. However, while the bottom boundary condition for wind (a null value at

79   the height of the aerodynamic level) allows defining a roughness length for momentum

80   exchange (z_{om}), the bottom value of the temperature profile, the aerodynamic temperature, is

81   generally unknown. One assumes usually that a roughness length z_{or} (improperly named

82   roughness length for thermal exchange) can be defined so that, at a height corresponding to

83   the displacement height plus z_{or}, the surface temperature of the vegetation can be considered

84   as representative of the aerodynamic temperature. The relationship between both roughness

85   lengths translates into what Matsushima names the "radiometric kB[-1]" (kB^{-1}_{radio}) which is

86   written as $z_{or} = z_{om} / e^{kB^{-1}_{radio}}$. In that case, the difference between the aerodynamic temperature

87   and the surface temperature in the sensible heat flux formulation is expressed by an additional

88   resistance (Lhomme et al., 1997):

89
$$H = \rho c_p \frac{T_{surf} - T_{air}}{r_{a,hm} + \frac{kB^{-1}_{radio}}{ku_*}} \Leftrightarrow H = \frac{\rho c_p k u_* \left(T_{surf} - T_{air} \right)}{\ln\left(\frac{z-d}{z_{or}} \right) - \Psi_h\left(\frac{z-d}{L} \right) + \Psi_h\left(\frac{z_{or}}{L} \right)}$$
(3)

90   where $r_{a,hm} = \left[\ln\left(\frac{z-d}{z_{om}} \right) - \Psi_h\left(\frac{z-d}{L} \right) + \Psi_h\left(\frac{z_{om}}{L} \right) \right] / ku_*$ is the aerodynamic resistance for

91   heat exchange before kB[-1] correction, u_* the friction velocity, k the von Karman constant, z the

92   measurement height of the atmospheric forcing, d the displacement height, L the Monin

93   Obhukhov length and ψ_h the stability correction function for heat transfer given by Paulson

94   (1970). kB^{-1}_{radio} is derived according to the expected air temperature profile within the canopy

95   and expressed as a function of meteorological data, *LAI* and plant height. Amongst the well

96   known formulae, one can cite those from Blümel (Blümel, 1998), Massman (Massman, 1999

97   revisited by Su et al., 2001) and Lhomme (Lhomme et al., 2000).

98   Other authors have proposed a somewhat simpler, and easier to interpret, formulation of the

99   relationship between T_{surf} and T_{aero}, called the "β function", originally proposed by Chehbouni

100   et al. (1997). β is expressed solely in terms of the temperatures, independently from wind

101   speed:

102   $$H = \rho c_p \beta \frac{T_{surf} - T_{air}}{r_{a,h}}$$ (4)

103   i.e.

104   $$\beta = \frac{T_{aero} - T_{air}}{T_{surf} - T_{air}}$$ (5)

105   Even for isothermal surfaces, usually bare soils or very dense canopies, the aerodynamic

106   temperature can be slightly different from the surface temperature, because the diffusion

107   process for heat transfer adds to the convective exchange of air. There is therefore a difference

108   between the effective eddy diffusivities for momentum and heat exchange, which can be again

109   translated into an excess resistance function of an "aerodynamic kB^{-1}" or kB^{-1}_{aero}. The

110   available kB^{-1} formulae, derived either empirically or from scalar and flux theoretical profiles

111   in the canopy, account for both aspects: the difference between the aerodynamic and the

112   radiometric temperatures (radiometric kB^{-1}) and the difference between momentum and heat

113   exchange diffusion processes (aerodynamic kB^{-1}). For isothermal surfaces, one can expect

114   that there is no radiometric component within the combined (radiometric and aerodynamic)

115   kB^{-1}. The combined kB^{-1} retrieved for those surfaces from observations by solving for kB^{-1}_{radio}

116   in Eq. 3 is fairly low (within the range 0-5 according to Verhoef et al. 1997, Massman 1999

117   and Yang et al., 2008). For strongly non-isothermal situations, which correspond in general to

118   intermediate *LAI* values (LAI in the range 0.5-2), the combined kB[-1] is much higher (in the

119   range 10-30 according to the same authors). One can thus assume that kB^{-1}_{aero} is usually

120   smaller than kB^{-1}_{radio} for all LAI values, or that the difference between the surface temperature

121   and the aerodynamic temperature will have on average a much larger impact on the sensible

122   heat flux than the difference between the diffusion processes for heat and momentum at the

123   vicinity of the canopy. While the radiometric kB^{-1} do not discriminate between both

124   differences (the difference between the aerodynamic and surface temperatures and the

125   difference between the roughness lengths for momentum z_{om} and heat exchange z_{oh}), the β

126   function allows us to separate both aspects and keep the difference between z_{oh} and z_{om} in the

127   formulation of the aerodynamic resistance : $z_{oh} = z_{om} \Big/ e^{kB^{-1}_{aero}}$ and $r_{a,h} = r_{a,hm} + kB^{-1}_{aero} \Big/ ku_*$.

128   Consequently, Eq. 4 can be rewritten as:

129    $$H = \rho c_p \beta \frac{T_{surf} - T_{air}}{r_{a,hm} + \dfrac{kB^{-1}_{aero}}{ku_*}} \Leftrightarrow H = \frac{\rho c_p \beta ku_* \left(T_{surf} - T_{air}\right)}{\ln\left(\dfrac{z-d}{z_{oh}}\right) - \Psi_h\left(\dfrac{z-d}{L}\right) + \Psi_h\left(\dfrac{z_{oh}}{L}\right)} \tag{6}$$

130   One must also note that the published values of the radiometric kB^{-1} (kB^{-1}_{radio}) (Matsushima,

131   2005) according to Lhomme, Blümel and Massman/Su (see formulations in Table 1) can be

132   converted into β values by combining equations 3 and 6:

133    $$\beta = \frac{r_{a,hm} + \dfrac{kB^{-1}_{aero}}{ku_*}}{r_{a,hm} + \dfrac{kB^{-1}_{radio}}{ku_*}} \tag{7}$$

134   The Blümel and Massman/Su formulations depend on Leaf Area Index *LAI* through, mostly,

135   the fraction cover (f_c) of the canopy and on two parameters difficult to assess, the component

136   aerodynamic kB^{-1} for bare soil and for vegetation canopy, respectively (See Table 1). Since

137   for bare soil and full cover conditions there is no large difference between both temperatures,

138   the β function is fairly easily interpreted: β values are close to 1 for those bare soil and full

139   cover conditions, that is, more generally, for all homogeneous isothermal surfaces, while for

140   sparse vegetation β decreases. In those conditions, the soil temperature has a large impact on

141   the radiative surface temperature whereas the aerodynamic temperature remains closer to a

142   mix of air and vegetation temperatures and is less influenced by the soil temperature. Since

143   the soil temperature around midday is generally higher than the vegetation temperature, the

144   observed radiative surface temperature is often larger than the aerodynamic temperature

145   around that time. Factors influencing β and kB^{-1}_{radio} include LAI and other plant geometrical

146   features such as height and fraction cover, friction velocity, time of the day, solar radiation etc.

147   However, most studies agree on the fact that LAI is by far the main driving factor, at least for

148   agricultural canopies for which the turbid medium (random leaf dispersion, Myneni et al.,

149   1989) and permeable-rough transfer hypotheses are valid (Kustas et al., 2007; Verhoef, 1997).

150   This is further confirmed by dual-source land surface models which predict the aerodynamic

151   temperature through the classical dual-source approach (Shuttleworth and Wallace, 1985).

152   The evolution of β as a function of *LAI* is presented in Figure 1a as obtained from an

153   uncalibrated run of the ICARE SVAT model (Gentine et al., 2007) for the B124 wheat site in

154   Morocco (see below). One can observe that the simulated shape of the $\beta(LAI)$ relationship

155   decreases sharply from 1 when LAI increase to about 1, and increases more slowly for higher

156   LAI, tending again to 1 for LAI well above 3. The lognormal distribution function is therefore

157   a good candidate to represent the evolution of 1-$\beta(LAI)$ for the whole range of LAI values.

158   Consequently, we propose the following empirical relationship for β :

159    $$\beta = 1 - \frac{a}{LAI * b * \sqrt{2\pi}} \cdot e^{-\frac{(\ln(LAI)-c)^2}{2*b^2}} \qquad (8)$$

160   where *a*, *b* and *c* are empirical coefficients that need to be calibrated.

161   The objectives of the present paper are threefold:

162   1- to retrieve β variations with LAI from observations for nine experimental cultural

163   cycles where seasonal evolution of factors governing β is available and LAI values

164   range from 0 to well above 2, and by doing so, assess the variability in shapes and

165   scales of the $\beta(LAI)$ relationship,

166   2- to compare several formulae of $\beta(LAI)$, including the new one (Eq. 8; hereafter

167   referred to as the Boulet et al. expression) , against observed trends and

168   3- to compare the performances of the various formulae in computing sensible heat flux

169   from observed in-situ radiometric surface temperature.

170   The new Boulet et al. expression (Eq. 8) of the β function will be first calibrated over the

171   values of β derived from experimental data acquired on the B124 wheat site in Morocco (i.e.

172   values will be obtained for a, b and c in Eq. 8). Next, it will be tested on data acquired over

173   eight other crop cycles in South of France.

174   2. STUDY SITES

175   All study sites cover an entire agricultural season, from bare soil to harvest.

176   The first dataset was collected at the B124 site (31.67250°N, 7.59597°W) in the R3 irrigation

177   perimeter in the Haouz semi-arid plain in Morocco during the SudMed project (Chehbouni et

178   al., 2008, Duchemin et al., 2006) in 2004. The climate is semi-arid with an average annual

179   rainfall of the order of 150 mm. The chosen field (number B124) was cultivated with winter

180   wheat and its size (4 ha) exceeded the basic fetch requirements. LAI and vegetation height

181   ranged from 0 in January to 4.5 and 0.8m at maximum development (April), respectively (See

182   Boulet et al., 2007, for more information on this dataset).

183 The second dataset was collected over two cultivated plots, Auradé (43°54'97''N,

184 01°10'61''E) and Lamasquère (43°49'65''N, 01°23'79''E), separated by 12 km and located

185 near Toulouse (South West France). The climate is Mediterranean with an average annual

186 rainfall of the order of 620 mm for both 2006 and 2007, which can be considered as "average"

187 years as far as rainfall is concerned. The Auradé plot was cultivated with winter wheat

188 (Triticum aestivum L., maximum LAI: 3.8, maximum height: 0.68 m) from Oct-2005 to Jun-

189 2006 and with sunflower (Helianthus annuus L., maximum LAI: 2.0, maximum height: 1.27

190 m) from Apr-2007 to Sept-2007. The Lamasquère plot was cultivated with maize (Zea mays

191 L., maximum LAI: 3.3, maximum height: 2.3 m) used for silaging from May-2006 to Aug-

192 2006 and with winter wheat from Oct-2006 to Jul-2007 (maximum LAI: 4.5, maximum

193 height: 0.83 m). This site was irrigated in 2006 when maize was cultivated. For a complete

194 description of the site characteristics, management practices, biomass inventories, vegetation

195 area measurements, instrumentation setup and fluxes calculation procedures see Beziat et al.

196 (2009).

197 The third dataset was acquired on the Avignon 'Flux and Remote Sensing Observation Site',

198 located in Provence, in southeastern France (N 43,92°; E 4,88°; altitude 32 m). The region is

199 also characterized by a typical Mediterranean climate (annual climatic mean of 14° C for

200 temperature and of 680 mm for precipitation). However, during the 2004-2007 period the

201 yearly average for rain was 450 mm with a high variability (313 mm – 745 mm). During the

202 observational period the crop rotation was: Durum Wheat (Triticum durum, maximum LAI:

203 4.0, maximum height: 1.0 m) from January till June 2004, Peas (Pisum sativum, maximum

204 LAI: 2.9, maximum height: 0.43 m) from April till June 2005, Durum Wheat (maximum LAI:

205 5.5, maximum height: 0.75 m) again from November 2005 to June 2006 and Sorghum

206 (Sorghum bicolor, maximum LAI: 3.0, maximum height: 1.16 m) from May to August 2007.

207 Irrigation was applied in particular to the Sorghum and Peas crops.

208   All sites were equipped with a tower where standard meteorological forcing data were

209   acquired. Plant height and total (green and dry) Leaf Area Index were measured every month

210   or so by planimetry and hemispherical photography and the resulting values have been

211   interpolated to daily estimates. Half hourly sensible heat fluxes were measured with a

212   Campbell Sci. CSAT3 (B124, Auradé and Lamasquère sites) or a Young 81000 (Avignon site)

213   3D sonic anemometer. Auradé, Lamasquère and Avignon are part of the CarboEurope-IP

214   Regional Experiment (Dolman et al., 2006) and the CarboEurope-IP Ecosystem Component.

215   In that context, the data were used for analyzing CO_2 surface – atmosphere exchanges and

216   production for fields with a large interannual rotation of crop types (e.g. Kutsch et al., 2010).

217   For those sites, the Level 3 flux products (i.e. non gapfilled) were used.

218   For all sites, TIR data were acquired with a nadir looking 60° Field Of View Apogee IRTS-P

219   (R3 and SudOuest sites) or an Heitronics KT15 (Avignon site) Infra Red Thermoradiometer

220   and a Kipp and Zonen CNR1 hemispherical pyrgeometer. For the R3 B124 site, the IRTS-P

221   device has been calibrated using an Everest black body during the experiment and prior to the

222   experiment in a laboratory with an adjustable ambient temperature. Retrieved surface

223   temperatures from both instruments showed a bias of 0.21°C and a Root Mean Square

224   Difference of 1.16°C for instantaneous values at midday. Surface temperature estimates from

225   the CNR1 were used for this site. The KT15 instrument in Avignon was calibrated in the

226   laboratory calibration facilities by looking at a black-body at various temperatures. An error

227   analysis of the whole calibration-measurement chain gave an accuracy of 0.4°C.

228   3. PROCEDURE FOR RETRIEVING β

229   For all sites, retrieved β function values estimated from observations, β_{obs}, are derived from

230   measured sensible heat H_{obs}, surface temperature $T_{surf,obs}$ and meteorological data such as air

231 temperature $T_{air,obs}$ that influence the aerodynamic resistance $r_{a,obs}$ calculated using the Monin-

232 Obukhov Similarity Theory:

$$\beta_{obs} = \frac{r_{a,obs} H_{obs}}{\rho c_p \left(T_{surf,obs} - T_{air,obs} \right)} \tag{9}$$

234 where $r_{a,obs}$ is computed using measured friction velocity values. Therefore:

$$r_{a,obs} = \frac{\left[\ln\left(\frac{z-d}{z_{om}/e^{kB^{-1}_{aero}}} \right) - \Psi_h\left(\frac{z-d}{L_{obs}} \right) + \Psi_h\left(\frac{z_{om}/e^{kB^{-1}_{aero}}}{L_{obs}} \right) \right]}{ku_{*,obs}} \tag{10}$$

236 Where L_{obs} is the Monin Obhukhov length estimated from observations and $u_{*,obs}$ the

237 measured friction velocity.

238 In order to assess either β (from Eq. 8) or the radiometric kB^{-1}, one would need in theory a

239 full determination of the three main variables: z_{om}, d and $z_{oh}=z_{om}/exp(kB^{-1}_{aero})$. This is feasible

240 from single-direction radiometric measurements only for bare soil and full cover conditions

241 since in the latter cases aerodynamic and surface temperatures can be considered as identical

242 ($kB^{-1}_{radio}=kB^{-1}_{aero}$ or $\beta=1$). For all other conditions, z_{om} and z_{oh} can be retrieved from u_* and

243 scale temperature $T_* = H/\rho c_p u_*$ measurements only if T_{aero} is already well known, i.e. if the

244 radiometric kB^{-1} is known. As a consequence, we decided to use the following expressions

245 from Colaizzi et al. (2004) and Pereira et al. (1999) to compute z_{om} and d:

$$d = h\left(1 - 2\frac{1 - e^{-0.5 LAI}}{LAI} \right) \tag{11}$$

$$z_{om} = he^{-0.5 LAI}\left(1 - e^{-0.5 LAI} \right) \tag{12}$$

248 where h is the vegetation height.

249   Given d, one can derive an estimated roughness length $z_{om,obs}$ according to:

250       $$z_{om,obs} \cong (z - d) e^{-\frac{ku_{obs}}{u_{*,obs}} - \Psi_m\left(\frac{z-d}{L_{obs}}\right)}$$ (13)

251   where u_{obs} is the wind velocity at reference height. We checked that z_{om} values obtained by Eq.

252   12 are valid, i.e. consistent with $z_{om,obs}$ obtained by Eq. 13 (not shown).

253   We choose emissivity values (between 0.96 for a bare soil and 0.98 for a vegetation at full

254   cover) for the computation of surface temperature from brightness temperature measurements

255   and a null value for the aerodynamic kB^{-1} ($e^{kB^{-1}_{aero}} = 1$) so that retrieved β_{obs} values deduced

256   from Eq. 9 tend to one for very high LAI values (and therefore retrieved radiometric kB^{-1} will

257   be close to zero when solving for kB^{-1}_{radio} in Eq. 7), which is what is expected for full cover

258   conditions. Values close to zero for the radiometric kB^{-1} can be found elsewhere in the

259   literature, notably for quasi-bare soils, as suggested both by Massman (1999)(in his paper the

260   radiometric kB^{-1} tends to 0 when LAI=0 or LAI>>2) and Matsushima (2005) in spite of the

261   large Reynolds numbers encountered in typical agricultural bare soil situations. Of course,

262   overall retrieval results and performances depend strongly on the accuracy of the surface

263   temperature measurements, and this should be investigated further, but it is beyond the scope

264   of this short paper. On the other hand, neglecting kB^{-1}_{aero} also leads to the best performance in

265   estimating sensible heat flux for all LAI values and all sites (not shown) which justifies our

266   choice. Yang et al. (2008) found values of the aerodynamic kB^{-1} for bare soils between -5 (at

267   night) and 5 (around midday) with an average between 0 and 5 for positive values of observed

268   sensible heat flux. In what follows, β_{obs} values are computed for unstable conditions selected

269   on the basis of the following rules: $H_{obs}>0$, total incoming solar radiation >10 W/m^2 and

270       $T_{surf,obs}>T_{air,obs}$. Retrieved β_{obs} values from Eq. 9 between -1 and 2 are kept since they

271   represent realistic values of the temperature profile within the canopy and the soil/vegetation

272   mixed/ensemble contribution (this represents >85% of all H_{obs} data between 10am and 4pm

273   for the R3 B124 dataset).

274

275   4. RESULTS FOR WHEAT AND OTHER CROPS

276   4.1. Results for the R3 B124 wheat site (calibration of Eq. 8)

277   Median values of the scattered β_{obs} values are shown for each 0.5 LAI interval on Figure 1a,

278   together with β values simulated by the ICARE dual source SVAT model applied to the R3

279   B124 site without any calibration of ICARE, β values interpolated along LAI values from

280   individual β estimates proposed by Matsushima (2005) as well as β values obtained from Eq.

281   8 (also referred to as "Boulet et al."), whose a, b and c coefficients are manually adjusted to

282   fit β_{obs} values: a=1.7, b=c=0.8. Error bars are shown for observed β values by assuming for

283   Eq. 9 an error of 1K on $T_{surf,obs}$, 10% on $u_{*,obs}$ and z_{om} and 10 W/m^2 on H_{obs}.

284   β values converted from Eq. 7 based on the three kB^{-1} expressions proposed by Blümel,

285   Massman/Su and Lhomme are also shown in Figure 1a.

286   For well developed vegetation (say, above LAI=2), all expressions are fairly similar or at least

287   show consistent trends. For sparse and less developed vegetation however (LAI<1, figure 1b,

288   where all single "observed" retrieved β values are shown instead of median values), there is a

289   large difference between the Blümel, Massman/Su, Matsushima and Lhomme expressions, on

290   the one hand, and ICARE outputs, the proposed Boulet et al. expression and the observations,

291   on the other hand. Note that ICARE has not been calibrated to the R3 dataset, therefore only

292   the overall shape of the $\beta(LAI)$ relationship is being used to introduce Eq. 8, not its particular

293   scaling/extent properties (a, b and c values) which are derived from β_{obs} values. The

294   differences between both groups of estimates could be explained by the fact that for the

Blümel, Massman/Su and Lhomme expressions, when *LAI* tends towards 0, the turbulent

behavior gives more weight to the aerodynamic kB^{-1} for a bare soil according to Brutsaert

(1982) which tends to underestimate β (Yang et al., 2008). As a consequence, the sensible heat

flux simulated from the different formulae for β using the observed surface temperature

(Equation 4), as one would do with remote sensing observations, is closer to the observed

sensible heat for the proposed new formula compared to the three previous ones, especially

for quasi bare soil (LAI<0.5) conditions. This appears clearly on Figure 2, where the resulting

sensible heat flux is estimated by using the various formulae for kB^{-1} or β together with Eqs. 3

and 4 respectively are shown. On this figure, the sensible heat flux values away from the 1:1

line for the Blümel, Massman/Su, Matsushima and Lhomme expressions correspond to those

conditions (LAI<0.5).

One must note that if we ignore the β correction (i.e. by setting β=1 or $kB^{-1}_{radio} = 0$ for all *LAI*

values) the root mean square error RMSE between observed and simulated sensible heat

fluxes increases from ~50 to ~80 W/m^2 for the entire season. This confirms that taking into

account the difference between the aerodynamic and the surface temperature is crucial to

deriving accurate turbulent heat fluxes.

4.2. Results for 8 agricultural seasons in Southern France (validation).

The different expressions for β are tested for the 3 other sites, i.e. 8 growing seasons and

different vegetation and climate conditions. We assume that the adjusted values of the three

parameters for the new formulation (*a*=1.7, *b*=*c*=0.8) can provide a good estimate of $\beta(LAI)$

fluctuations for agricultural areas. These values are therefore kept for all sites. Results are

shown in Figure 3. Surface temperature is estimated either from the hemispherical (CNR1) or

the directional (KT15, IRTS-P) sensors, as specified in the captions. In general, the

hemispherical device produces a $\beta(LAI)$ relationship closer to the expected concave-up shape.

319   For all sites, the new formulation (Eq. 8) fits fairly well the observed β, except for

320   Lamasquère in 2007, which means that the calibrated values for *a*, *b* and *c* at the R3 B124 site

321   are well suited for other sites. For sunflower, corn and sorghum however (Figure 3c, e and f),

322   the trend in $\beta(LAI)$ matches the observed trend but not the observed amplitude, and the value

323   for *a* should therefore be lower for those cover types. This might be due to the size and shape

324   of the leaves, or to the fact that these canopies show more defined geometrical features (rows,

325   preferred orientation of leaves and flowers…) and are less easily described by turbid medium

326   (for radiation) or permeable-rough interfacial layer (for turbulent transfer) theories. It is also

327   possible to adjust the *a*, *b* and *c* values to fit more closely the observed $\beta(LAI)$ curve. Adjusted

328   values for *a* range between 1.2 and 1.7, and adjusted values for *b* and *c* between 0.5 and 0.8,

329   but do not translate into a significant improvement in computing sensible heat flux.

330

331   Since the primary objective of the β formulation is to provide accurate estimates of sensible

332   heat fluxes from observed surface temperature, one needs to assess the resulting performance

333   in using the various β models to simulate *H*. However, it should be noted that the accuracy of

334   those estimates depends primarily on the precision on surface and air temperatures, wind

335   speed, roughness length and the validity of the Monin-Obukhov Similarity Theory.

336   The performance in estimating sensible heat flux for all sites was assessed using the various β

337   formulations. Results are shown in Table 2. All formulations perform well for some sites and

338   much less well for others, but the new formulation shows comparable or better performances

339   than the others as indicated by the small number of RMSE values above 60 W/m^2 as well as

340   the high number of values under 50 W/m^2. The simplicity of this formulation and the fact that

341   it allows tuning at least one parameter (*a*) to provide realistic values of the difference between

342   the aerodynamic and the surface temperature are two advantages of this formulation. One

343   must note that, for both the Blümel and Massman formulations, the poorly known parameters

344   corresponding to the aerodynamic kB^{-1} for bare soil and full cover (which are found to be

345   close to zero here) can be adjusted as well as the parameter of the exponential decay of the

346   Beer Lambert law used to convert *LAI* to fraction cover, but in our simulations the results did

347   not prove to be very sensitive to the latter (not shown). Again, all formulations perform well

348   for canopies well described by the turbid medium theory, and less for plants with more

349   defined geometrical features or larger intercropping patterns (Barillot et al., 2011).

350

351   5. CONCLUSION

352   An empirical formulation of the difference between the aerodynamic and surface temperatures

353   as a function of Leaf Area Index has been proposed, which represents in a realistic way the

354   observed variations and leads to satisfactory performance in simulating the sensible heat flux

355   compared to other existing formulations. It should be noted though that the observed

356   variations in this difference were assessed using a null aerodynamic kB^{-1}, based on the fact

357   that the radiometric kB^{-1} should be close to zero (or β close to one) for bare soil and full cover

358   conditions. However, the assumption that the difference between the surface temperature and

359   the aerodynamic temperature will have on average a much larger impact on the sensible heat

360   than the difference between the diffusion processes for heat and momentum at the vicinity of

361   the canopy should be investigated more thoroughly since there is no agreement on what value

362   should be used for the aerodynamic kB^{-1} for bare soil, intermediate cover and fully covering

363   vegetation, respectively (Verhoef, 1997). In particular, there is no evidence whatsoever that

364   the roughness length for heat exchange should be a constant fraction of the roughness length

365   for momentum for all LAI values, as it is commonly assumed in SVAT models.

366

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REFERENCES

Barillot, R., Louarn, G., Escobar-Gutierrez, A.J., Huynh, P., Combes, D., 2011. How good is the turbid medium-based approach for accounting for light partitioning in contrasted grass-legume intercropping systems ? Ann. Bot., 108(6): 1013-1024.

Beziat, P., Ceschia, E. and Dedieu, G., 2009. Carbon balance of a three crop succession over two cropland sites in South West France. Agr. For. Met., 149(10): 1628-1645.

Blümel, K., 1998. Estimation of sensible heat flux from surface temperature wave and one-time-of-day air temperature observation. Boundary-Layer Meteorol., 86(2): 193-232.

Boulet, G., Chehbouni, A., Gentine, P., Duchemin, B., Ezzahar, J. and Hadria, R., 2007. Monitoring water stress using time series of observed to unstressed surface temperature difference. Agr. For. Met., 146(3-4): 159-172.

Brutsaert, W., 1982. Evaporation into the atmosphere. Environmental fluid mechanics. Kluwer Academic Publishers, Dordrecht / Boston / London, 299 pp.

395   Carlson, T.N., Taconet, O., Vidal, A., Gilles, R.R., Olioso, A. and Humes, K., 1995. an
396   overview of the workshop on thermal remote-sensing held at La-Londe-Les-Maures,
397   France, September 20-24, 1993. Agr. For. Met., 77(3-4): 141-151.

398   Chehbouni, A., LoSeen, D., Njoku, E.G., Lhomme, J.P., Monteny, B. and Kerr, Y.H., 1997.
399   Estimation of sensible heat flux over sparsely vegetated surfaces. J. Hydrol., 189(1-4):
400   855-868.

401   Chehbouni, A., Escadafal, R., Duchemin, B., Boulet, G., Simonneaux, V., Dedieu, G.,
402   Mougenot, B., Khabba, S., Kharrou, H., Maisongrande, P., Merlin, O., Chaponniere,
403   A., Ezzahar, J., Er-Raki, S., Hoedjes, J., Hadria, R., Abourida, A., Cheggour, A.,
404   Raibi, F., Boudhar, A., Benhadj, I., Hanich, L., Benkaddour, A., Guemouria, N.,
405   Chehbouni, A.H., Lahrouni, A., Olioso, A., Jacob, F., Williams, D.G. and Sobrino,
406   J.A., 2008. An integrated modelling and remote sensing approach for hydrological
407   study in arid and semi-arid regions: the SUDMED programme. Int. J. Remote Sens.,
408   29(17-18): 5161-5181.

409   Cleugh, H.A., Leuning, R., Mu, Q.Z. and Running, S.W., 2007. Regional evaporation
410   estimates from flux tower and MODIS satellite data. Remote Sens. Environ., 106(3):
411   285-304.

412   Colaizzi, P.D., Evett, S.R., Howell, T.A. and Tolk, J.A., 2004. Comparison of aerodynamic
413   and radiometric surface temperature using precision weighing lysimeters. In: W. Gao
414   and D.R. Shaw (Editors), Remote Sensing and Modeling of Ecosystems for
415   Sustainability. Proceedings of the Society of Photo-Optical Instrumentation Engineers
416   (Spie). Spie-Int Soc Optical Engineering, Bellingham, pp. 215-229.

417   Courault, D., Seguin, B., Olioso, A., 2005. Review on estimation of evapotranspiration from
418   remote sensing data: From empirical to numerical modeling approaches. Irrig. Drain.
419   Syst., 19(3-4): 223-249.

420   Dolman, A.J., Noilhan, J., Durand, P., Sarrat, C., Brut, A., Piguet, B., Butet, A., Jarosz, N.,
421   Brunet, Y., Loustau, D., Lamaud, E., Tolk, L., Ronda, R., Miglietta, F., Gioli, B.,
422   Magliulo, V., Esposito, M., Gerbig, C., Korner, S., Glademard, R., Ramonet, M.,
423   Ciais, P., Neininger, B., Hutjes, R.W.A., Elbers, J.A., Macatangay, R., Schrems, O.,
424   Perez-Landa, G., Sanz, M.J., Scholz, Y., Facon, G., Ceschia, E. and Beziat, P., 2006.
425   The CarboEurope regional experiment strategy. Bull. Am. Meteorol. Soc., 87(10):
426   1367-1379.

427   Duchemin, B., Hadria, R., Erraki, S., Boulet, G., Maisongrande, P., Chehbouni, A., Escadafal,
428   R., Ezzahar, J., Hoedjes, J.C.B., Kharrou, M.H., Khabba, S., Mougenot, B., Olioso,
429   A., Rodriguez, J.C., Simonneaux, V., 2006. Monitoring wheat phenology and
430   irrigation in Central Morocco: On the use of relationships between evapotranspiration,
431   crops coefficients, leaf area index and remotely-sensed vegetation indices.
432   Agricultural Water Management, 79(1): 1-27.

433 Gentine, P., Entekhabi, D., Chehbouni, A., Boulet, G. and Duchemin, B., 2007. Analysis of
434   evaporative fraction diurnal behaviour. Agr. For. Met., 143(1-2): 13-29.

435 Gowda, P., Chavez, J., Colaizzi, P., Evett, S., Howell, T. and Tolk, J., 2008. ET mapping for
436   agricultural water management: present status and challenges. Irrig. Sci., 26(3): 223-
437   237.

438 Jia, L., Li, Z.L., Menenti, M., Su, Z., Verhoef, W., Wan, Z., 2003. A practical algorithm to
439   infer soil and foliage component temperatures from bi-angular ATSR-2 data.
440   International Journal of Remote Sensing, 24(23): 4739-4760.

441 Kalma, J.D. and Jupp, D.L.B., 1990. Estimating evaporation from pasture using infrared
442   thermometry: evaluation of a one-layer resistance model. Agr. For. Met., 51(3-4): 223-
443   246.

444 Kalma, J.D., McVicar, T.R. and McCabe, M.F., 2008. Estimating Land Surface Evaporation:
445   A Review of Methods Using Remotely Sensed Surface Temperature Data. Surv.
446   Geophys., 29(4-5): 421-469.

447 Kustas, W. and Anderson, M., 2009. Advances in thermal infrared remote sensing for land
448   surface modeling. Agr. For. Met., 149(12): 2071-2081.

449 Kustas, W.P., Anderson, M.C., Norman, J.M. and Li, F.Q., 2007. Utility of radiometric-
450   aerodynamic temperature relations for heat flux estimation. Boundary-Layer
451   Meteorol., 122(1): 167-187.

452 Kutsch, W.L., Aubinet, M., Buchmann, N., Smith, P., Osborne, B., Eugster, W., Wattenbach,
453   M., Schrumpf, M., Schulze, E.D., Tomelleri, E., Ceschia, E., Bernhofer, C., Beziat, P.,
454   Carrara, A., Di Tommasi, P., Grunwald, T., Jones, M., Magliulo, V., Marloie, O.,
455   Moureaux, C., Olioso, A., Sanz, M.J., Saunders, M., Sogaard, H. and Ziegler, W.,
456   2010. The net biome production of full crop rotations in Europe. Agriculture
457   Ecosystems & Environment, 139(3): 336-345.

458 Lhomme, J.P., Chehbouni, A. and Monteny, B., 2000. Sensible heat flux-radiometric surface
459   temperature relationship over sparse vegetation: Parameterizing B-1. Boundary-Layer
460   Meteorol., 97(3): 431-457.

461 Lhomme, J.P., Troufleau, D., Monteny, B., Chehbouni, A. and Bauduin, S., 1997. Sensible
462   heat flux and radiometric surface temperature over sparse Sahelian vegetation .2.
463   Model for the kB(-1) parameter. J. Hydrol., 189(1-4): 839-854.

464 Massman, W.J., 1999. A model study of kB(H)(-1) for vegetated surfaces using 'localized
465   near-field' Lagrangian theory. J. Hydrol., 223(1-2): 27-43.

466 Matsushima, D., 2005. Relations between aerodynamic parameters of heat transfer and
467   thermal-infrared thermometry in the bulk surface formulation. J. Meteorol. Soc. Japan,
468   83(3): 373-389.

469 Myneni, R.B., Ross, J., Asrar, G., 1989. A review on the theory of photon transport in leaf
470   canopies. Agricultural and Forest Meteorology, 45(1-2): 1-153.

471 Norman, J.M. and Becker, F., 1995. Terminology in thermal infrared remote-sensing of
472   natural surfaces. Agr. For. Met., 77(3-4): 153-166.

473 Olioso, A., Inoue, Y., Ortega-Farias, S., Demarty, J., Wigneron, J.P., Braud, I., Jacob, F.,
474   Lecharpentier, P., Ottlé, C., Calvet, J.C. and Brisson, N., 2005. Future directions for
475   advanced evapotranspiration modeling: Assimilation of remote sensing data into crop
476   simulation models and SVAT models. Irrig. Drain. Sys., 19(3): 377-412.

477 Paulson, C.A., 1970. The Mathematical Representation of Wind Speed and Temperature
478   Profiles in the Unstable Atmospheric Surface Layer. J. Appl. Meteorol., 9(6): 857-861.

479 Pereira, L.S., Perrier, A., Allen, R.G. and Alves, I., 1999. Evapotranspiration: Concepts and
480   future trends. Irrig. Drain. Syst., 125(2): 45-51.

481 Shuttleworth, W.J., and Wallace, J.S., 1985. Evaporation from sparse crops - an energy
482   combination theory. Q. J. R. Meteorol. Soc., 111: 839-855.

483 Stewart, J.B., Kustas, W.P., Humes, K.S., Nichols, W.D., Moran, M.S. and Debruin, H.A.R.,
484   1994. Sensible heat-flux radiometric surface-temperature relationship for 8 semiarid
485   areas. J. Appl. Meteorol., 33(9): 1110-1117.

486 Su, Z., Schmugge, T., Kustas, W.P. and Massman, W.J., 2001. An evaluation of two models
487   for estimation of the roughness height for heat transfer between the land surface and
488   the atmosphere. J. Appl. Meteorol., 40(11): 1933-1951.

489 Venturini, V., Islam, S. and Rodriguez, L., 2008. Estimation of evaporative fraction and
490   evapotranspiration from MODIS products using a complementary based model.
491   Remote Sens. Environ., 112(1): 132-141.

492 Verhoef, A., de Bruin, H. A. R., and van den Hurk, B. J. J. M., 1997. Some Practical Notes on
493   the Parameter kB^{-1} for Sparse Vegetation. J. Appl. Meteorol., 36: 560-572.

494 Yang, K., Koike, T., Ishikawa, H., Kim, J., Li, X., Liu, H.Z., Liu, S.M., Ma, Y.M. and Wang,
495   J.M., 2008. Turbulent flux transfer over bare-soil surfaces: Characteristics and
496   parameterization. J. Appl. Meteorol. Climatol., 47(1): 276-290.