SOIL Discuss., 2, 737–765, 2015
www.soil-discuss.net/2/737/2015/
doi:10.5194/soild-2-737-2015

SOIL Discussions Open Access

Impact of gravels and organic matter on the thermal properties of grassland soils in southern France

J.-C. Calvet, N. Fritz, C. Berne, B. Piguet, W. Maurel, and C. Meurey

CNRM-GAME UMR 3589 (Météo-France, CNRS), Toulouse, France

Received: 20 May 2015 – Accepted: 10 June 2015 – Published: 26 June 2015

Correspondence to: J.-C. Calvet (jean-christophe.calvet@meteo.fr)

Published by Copernicus Publications on behalf of the European Geosciences Union.

Abstract

Soil moisture is the main driver of temporal changes in values of the soil thermal conductivity. The latter is a key variable in land surface models (LSMs) used in hydrometeorology, for the simulation of the vertical profile of soil temperature in relation to soil moisture. Shortcomings in soil thermal conductivity models tend to limit the impact of improving the simulation of soil moisture in LSMs. Models of the thermal conductivity of soils are affected by uncertainties, especially in the representation of the impact of soil properties such as the volumetric fraction of quartz (q), soil organic matter, and gravels. As soil organic matter and gravels are often neglected in LSMs, the soil thermal conductivity models used in most LSMs represent the mineral fine earth, only. Moreover, there is no map of q and it is often assumed that this quantity is equal to the volumetric fraction of sand. In this study, q values are derived by reverse modelling from the continuous soil moisture and soil temperature sub-hourly observations of the Soil Moisture Observing System – Meteorological Automatic Network Integrated Application (SMOSMANIA) network at 21 grassland sites in southern France, from 2008 to 2015. The soil temperature observations are used to retrieve the soil thermal diffusivity (D_h) at a depth of 0.10 m in unfrozen conditions, solving the thermal diffusion equation. The soil moisture and D_h values are then used together with the measured soil properties to retrieve soil thermal conductivity (λ) values. For ten sites, the obtained λ value at saturation (λ_{sat}) cannot be retrieved or is lower than the value corresponding to a null value of q, probably in relation to a high density of grass roots at these sites or to the presence of stones. For the remaining eleven sites, q is negatively correlated with the volumetric fraction of solids other than sand. The impact of neglecting gravels and organic matter on λ_{sat} is assessed. It is shown that these factors have a major impact on λ_{sat}.

SOILD

2, 737–765, 2015

1 Introduction

Soil thermal properties are characterized by two key variables: the soil volumetric heat capacity (C_h), and the soil thermal conductivity (λ), in $\mathrm{J\,m^{-3}\,K^{-1}}$ and $\mathrm{W\,m^{-1}\,K^{-1}}$, respectively. Provided the volumetric fractions of moisture, minerals and organic matter are known, C_h can be calculated easily. On the other hand, the estimation of λ relies on empirical models and is affected by uncertainties (Peters-Lidard et al., 1998 ; Tarnawski et al., 2012). The construction and the verification of the λ models is not easy as λ is difficult to measure directly in situ and is often measured in the lab on perturbed soil samples (Abu-Hamdeh et al., 2000). Moreover, for given soil moisture conditions, λ depends to a large extent on the fraction of soil minerals presenting high thermal conductivities such as quartz, hematite, dolomite or pyrite (Côté and Conrad, 2005). At mid-latitudes, quartz is the main driver of λ. The fraction of quartz is generally unknown as it can only be measured using X-ray diffraction or X-ray fluorescence techniques, which are difficult to implement (Schönenberger et al., 2012).

Today, most of the Land Surface Models (LSMs) used in meteorology and hydrometeorology simulate λ following the approach proposed by Peters-Lidard et al. (1998). This approach consists of an updated version of the Johansen (1975) model, and assumes that the volumetric fraction of quartz (q) is equal to the volumetric fraction of sand (f_{sand}). This is a strong assumption, as some sandy soils (e.g. calcareous sands) may contain little quartz, and as quartz may be found in the silt and clay fractions of the soil minerals. Moreover, soil organic matter (SOM) and gravels are often neglected in LSMs, and the λ models used in most LSMs represent the mineral fine earth, only. Yang et al. (2005) and Chen et al. (2012) have shown the importance of accounting for SOM and gravels in λ models for organic top soil layers of grasslands of the Tibetan plateau.

In this study, an attempt is made to use routine automatic soil moisture and soil temperature sub-hourly measurements to retrieve instantaneous λ values at 21 weather stations of the Soil Moisture Observing System – Meteorological Automatic Network

SOILD
2, 737–765, 2015

Integrated Application (SMOSMANIA) network (Calvet et al., 2007) in southern France, at a depth of 0.10 m. The response of λ to soil moisture is investigated and the feasibility of modelling the λ value at saturation (λ_{sat}) with or without using SOM and gravel fraction observations is assessed. The q values are retrieved by reverse modelling.

The field data and the method to retrieve λ values are presented in Sect. 2. The λ and q retrievals are presented in Sect. 3 together with a sensitivity analysis of λ_{sat} to SOM and gravel fractions. Finally, the results are discussed in Sect. 4, and the main conclusions are summarized in Sect. 5.

2 Data and methods

2.1 The SMOSMANIA data

The SMOSMANIA soil moisture network was developed by Calvet et al. (2007) in southern France in order to validate satellite-derived soil moisture products (Parrens et al., 2012), assess land surface models used in hydrological models (Draper et al., 2011) and in meteorological models (Albergel et al., 2010), and monitor the impact of climate change on water resources and droughts. The station network forms a transect between the Atlantic coast and the Mediterranean sea (Fig. 1). It consists of pre-existing automatic weather stations operated by Meteo-France, upgraded with four soil moisture probes at four depths: 0.05, 0.10, 0.20, and 0.30 m. In general, the stations are located on former cultivated fields and consist of grasslands. Soil properties were measured at each stations using soil samples collected during the installation of the probes. The 21 stations cover a very large range of soil texture characteristics (Fig. 2). At the same time, Fig. 2 shows that soil texture does not vary much with depth (from 0.05 to 0.20 m) at a given station. Other properties such as the gravimetric fraction of the Soil Organic Matter (SOM) and of gravels were determined from the soil samples. In addition, the bulk dry density of the soil (ρ_d) was measured using unperturbed oven-dried soil samples collected using metal cylinders of known volume (about 7×10^{-4} m^3).

SOILD
2, 737–765, 2015

Twelve SMOSMANIA stations were activated in 2006 in southwestern France. In 2008, nine more stations were installed along the Mediterranean coast, and the whole network (21 stations) was gradually equipped with temperature sensors at the same depths as soil moisture probes. The soil moisture and soil temperature probes consisted of Thetaprobe ML2X and PT100 sensors, respectively.

The ThetaProbe sensors provide a voltage signal in units of V. In order to convert the voltage signal into volumetric soil moisture content ($m^3\,m^{-3}$), site-specific calibration curves were developed using in situ gravimetric soil samples for all stations, and for all depths (Albergel et al., 2008). In this study, the calibration was revised in order to avoid spurious high soil moisture values during intense precipitation events. Logistics curves were used instead of exponential curves in the previous version of the data set.

The soil temperature observations are recorded with a resolution of 0.1 °C.

The observations from the 48 soil moisture probes and from the 48 temperature probes are automatically recorded every 12 min. The data are available to the research community through the International Soil Moisture Network web site (https://ismn.geo.tuwien.ac.at/).

Figure 3 shows soil temperature time series at the Saint-Félix-de-Lauragais (SFL) station on 23 February 2015. The impact of recording temperature with a resolution of 0.1 °C is clearly visible at all depths as this causes a levelling of the curves.

In this study, sub-hourly measurements of soil temperature and soil moisture at a depth of 0.10 m are used, together with soil temperature measurements at 0.05 and 0.20 m, from 1 January 2008 to 28 February 2015.

2.2 Soil characteristics

The porosity values at a depth of 0.10 m are listed in Table 1 together with gravimetric and volumetric fractions of soil particle-size ranges (sand, clay, silt, gravel) and SOM. The porosity, or soil volumetric moisture at saturation (θ_{sat}), is derived from the bulk dry density ρ_d, together with soil texture and soil organic matter observations as:

$$\theta_\mathrm{sat} = 1 - \rho_\mathrm{d}\left[\frac{m_\mathrm{sand} + m_\mathrm{clay} + m_\mathrm{silt} + m_\mathrm{gravel}}{\rho_\mathrm{min}} + \frac{m_\mathrm{SOM}}{\rho_\mathrm{SOM}}\right]$$

or

$$\theta_\mathrm{sat} = 1 - f_\mathrm{sand} - f_\mathrm{clay} - f_\mathrm{silt} - f_\mathrm{gravel} - f_\mathrm{SOM}, \tag{1}$$

where m_x (f_x) represents the gravimetric (volumetric) fraction of the soil component x. The f_x values are derived from the measured gravimetric fractions, multiplied by the ratio of ρ_d observations to ρ_x, the density of each soil component x. Values of $\rho_\mathrm{SOM} = 1300\,\mathrm{kg\,m^{-3}}$ and $\rho_\mathrm{min} = 2660\,\mathrm{kg\,m^{-3}}$ are used for soil organic matter, and soil minerals, respectively.

2.3 Retrieval of soil thermal diffusivity

The soil thermal diffusivity (D_h) is expressed in $\mathrm{m^2\,s^{-1}}$ and is defined as:

$$D_h = \frac{\lambda}{C_\mathrm{h}} \tag{2}$$

In this study, a simple numerical method is used to retrieve instantaneous values of D_h at a depth of 0.10 m using three soil temperature observations at 0.05, 0.10 and 0.20 m, performed every 12 min, by solving the Fourier thermal diffusion equation. The latter can be written as:

$$C_\mathrm{h}\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda\frac{\partial T}{\partial z}\right). \tag{3}$$

In this study, given that soil properties are relatively homogeneous on the vertical (Sect. 2.1), values of D_h can be derived from the Fourier one-dimensional law:

$$\frac{\partial T}{\partial t} = D_h\frac{\partial^2 T}{\partial z^2} \tag{4}$$

SOILD
2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

Given that three soil temperatures T_i (i ranging from 1 to 3) are measured at depths $z_1 = -0.05$ m, $z_2 = -0.10$ m, and $z_3 = -0.20$ m, the soil diffusivity D_{hi} at $z_i = z_2 = -0.10$ m can be obtained by solving the one-dimensional heat equation, using a finite difference method based on the implicit Crank-Nicholson scheme. When three soil depths are considered, z_{i-1}, z_i, z_{i+1}, the change in soil temperature T_i at depth z_i, from time t_{n-1} to time t_n, within the time interval $\Delta t = t_n - t_{n-1}$ can be written as:

$$\frac{T_i^n - T_i^{n-1}}{\Delta t} = D_{hi}\left[\frac{1}{2}\left(\frac{\gamma_{i+1}^n - \gamma_i^n}{\Delta z_m}\right) + \frac{1}{2}\left(\frac{\gamma_{i+1}^{n-1} - \gamma_i^{n-1}}{\Delta z_m}\right)\right] \text{ with}$$
$$\gamma_i^n = \frac{T_i^n - T_{i-1}^n}{\Delta z_i}, \Delta z_m = \frac{\Delta z_i + \Delta z_{i+1}}{2}, \text{ and } \Delta z_i = z_i - z_{i-1}. \tag{5}$$

In this study, $\Delta z_i = -0.05$ m, $\Delta z_{i+1} = -0.10$ m, and a value of $\Delta t = 2880$ s (48 min) is used.

It is important to ensure that D_h retrievals are related to diffusion processes only and not to the transport of heat by water infiltration or evaporation (Parlange et al., 1998; Schelde et al., 1998). Therefore, only situations for which changes in soil moisture at all depths do not exceed 0.001 $m^3\,m^{-3}$ within the Δt time lag are considered.

2.4 From soil diffusivity to soil thermal conductivity

The observed soil properties and volumetric soil moisture are used to calculate the volumetric heat capacity C_h at a depth of 0.10 m. The C_h values, in units of $J\,m^{-3}\,K^{-1}$, are calculated as:

$$C_h = \theta C_{hwater} + f_{min} C_{hmin} + f_{SOM} C_{hSOM} \tag{6}$$

and values of $4.2 \times 10^6\,J\,m^{-3}\,K^{-1}$, $2.0 \times 10^6\,J\,m^{-3}\,K^{-1}$, and $2.5 \times 10^6\,J\,m^{-3}\,K^{-1}$, are used for C_{hwater}, C_{hmin}, C_{hSOM}, respectively.

The λ values at 0.10 m are then derived from the D_h and C_h estimates (Eq. 2).

SOILD
2, 737–765, 2015

2.5 Soil thermal conductivity model

In dry conditions, soils present low thermal conductivity values (λ_{dry}). Experimental evidence show that λ_{dry} is negatively correlated with porosity. For example, Lu et al. (2007) give:

$$\lambda_{dry} = 0.51 - 0.56 \times \theta_{sat} \quad (\text{in W m}^{-1}\,\text{K}^{-1}). \tag{7}$$

When soil pores are gradually filled with water, λ tends to increase towards a maximum value at saturation (λ_{sat}). Between dry and saturation conditions, λ is expressed as:

$$\lambda = \lambda_{dry} + K_e \left(\lambda_{sat} - \lambda_{dry}\right) \tag{8}$$

where, K_e is the Kersten number. The latter is related to the volumetric soil moisture, θ, i.e. to the degree of saturation (S_d). In this study, the formula recommended by Yang et al. (2005) is used:

$$K_e = \exp\left(k_T \left(1 - 1/S_d\right)\right),$$

with

$$k_T = 0.36 \text{ and } S_d = \theta \,/\, \theta_{sat}. \tag{9}$$

The geometric mean equation for λ_{sat} proposed by Johansen (1975) for the mineral components of the soil can be generalized to include the SOM thermal conductivity (Chen et al., 2012) as:

$$\ln(\lambda_{sat}) = q \ln\left(\lambda_q\right) + f_{other} \ln(\lambda_{other}) + \theta_{sat} \ln(\lambda_{water}) + f_{SOM} \ln(\lambda_{SOM}), \tag{10}$$

where q is the volumetric fraction of quartz, and $\lambda_q = 7.7\,\text{W m}^{-1}\,\text{K}^{-1}$, $\lambda_{other} = 2.0\,\text{W m}^{-1}\,\text{K}^{-1}$, $\lambda_{water} = 0.594\,\text{W m}^{-1}\,\text{K}^{-1}$, $\lambda_{SOM} = 0.25\,\text{W m}^{-1}\,\text{K}^{-1}$ are the thermal conductivities of quartz, soil minerals other than quartz, water and SOM, respectively. The volumetric fraction of soil minerals other than quartz is defined as:

$$f_{other} = 1 - q - \theta_{sat} - f_{SOM}. \tag{11}$$

Discussion Paper | Discussion Paper | Discussion Paper | Discussion Paper |

SOILD

2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

Title Page
Abstract Introduction
Conclusions References
Tables Figures
◂◂ ▸▸
◂ ▸
Back Close
Full Screen / Esc

Printer-friendly Version

Interactive Discussion

2.6 Reverse modelling

The λ_{sat} values are retrieved through reverse modelling using the λ model described above (Eqs. 7–11). The λ model is used to produce simulations of λ at the same soil moisture conditions as those encountered for the λ values derived from observations in Sect. 2.4. For a given station, a set of 401 λ simulations is produced for λ_{sat} ranging from 0 to 4 $W\,m^{-1}\,K^{-1}$, with a resolution of 0.01 $W\,m^{-1}\,K^{-1}$. The λ_{sat} retrieval corresponds to the λ simulation presenting the lowest root mean square difference (RMSD) value with respect to the observations. Only λ observations for S_d values higher than 0.4 are used, as the λ_{sat} retrievals are very sensitive to uncertainties in λ observations obtained in dry conditions. Finally, the q value is derived from the retrieved λ_{sat} by solving Eq. (10), provided at least twenty λ observations can be used. When negative values of q are obtained, a null value of q is imposed

3 Results

3.1 λ_{sat} and q retrievals

Figure 4 shows retrieved and modelled λ values vs. the observed degree of saturation of the soil, at a depth of 0.10 m, for contrasting retrieved values of λ_{sat} (2.79, 1.45, and 0.70 $W\,m^{-1}\,K^{-1}$) at the SBR, MNT, and PRD stations, respectively. All the obtained λ_{sat} and q retrievals are listed in Table 2, together with the λ RMSD values and the number of available λ observations. For six stations (CRD, PZN, MZN, VLV, MJN, and BRZ), the reverse modelling technique described in Sect. 2.6 cannot be implemented as not enough λ observations could be obtained for S_d values higher than 0.4. For the other 15 stations, λ_{sat} and q retrievals are obtained using 62 to 1939 λ observations. For the five stations (LHS, SVN, NBN, and PRD) presenting the lowest λ_{sat} retrievals, ranging between 0.52 and 1.11 $W\,m^{-1}\,K^{-1}$, null values of q are obtained. The λ model (Eqs. 7–

SOILD
2, 737–765, 2015

J.-C. Calvet et al.

11) is fully operative, with non-null values of q, for eleven stations: SBR, URG, PRG, CDM, MNT, SFL, MTM, LZC, LGC, BRN, CBR.

3.2 A pedotransfer function for quartz

The q retrievals can be used to assess the possibility to estimate q using other soil characteristics, which can be easily measured.

For the 11 stations with $q > 0$, a good correlation is found between q retrievals and f_{sand} ($r^2 = 0.66$, F test p value $= 0.0025$, RMSD $= 0.09\,\text{m}^3\,\text{m}^{-3}$). However, a better result is found considering the sum of the fractions of the other soil particles, and the modelled q values can be derived from the following pedotransfer function:

$$q_{\text{MOD}} = 0.70 - 1.075 \times \left(f_{\text{clay}} + f_{\text{silt}} + f_{\text{gravel}} + f_{\text{SOM}}\right)$$
$$\text{or}\;\; q_{\text{MOD}} = 0.70 - 1.075 \times (1 - \theta_{\text{sat}} - f_{\text{sand}}) \tag{12}$$

($r^2 = 0.78$, F test p value $= 0.0003$, RMSD $= 0.07\,\text{m}^3\,\text{m}^{-3}$). The values of q_{MOD} vs. q are shown in Fig. 5.

Modelled values of λ_{sat} (λ_{satMOD}) can be derived from q_{MOD} using Eq. (10) and the following r^2, RMSD, and mean bias scores are obtained for λ_{satMOD}, with respect to the λ_{sat} retrievals: 0.87, $0.15\,\text{W}\,\text{m}^{-1}\,\text{K}^{-1}$, and $-0.01\,\text{W}\,\text{m}^{-1}\,\text{K}^{-1}$, respectively (Table 3).

Finally, we investigated the possibility of estimating θ_{sat} from the soil characteristics listed in Table 1 and of deriving a statistical model for θ_{sat} (θ_{satMOD}). We found the following statistical relationship between θ_{satMOD}, m_{clay}, m_{silt}, and m_{SOM}:

$$\theta_{\text{satMOD}} = 0.456 - 0.0735\,\frac{m_{\text{clay}}}{m_{\text{silt}}} + 2.238\,m_{\text{SOM}} \tag{13}$$

($r^2 = 0.48$, F test p value $= 0.0027$, RMSD $= 0.036\,\text{m}^3\,\text{m}^{-3}$).

SOILD
2, 737–765, 2015

Volumetric fractions of soil components need to be consistent with θ_{satMOD} and can be calculated using the modelled bulk density values, derived from θ_{satMOD} as:

$$\rho_{dMOD} = \frac{1 - \theta_{satMOD}}{\frac{m_{sand}+m_{clay}+m_{silt}+m_{gravel}}{\rho_{min}} + \frac{m_{SOM}}{\rho_{SOM}}} . \tag{14}$$

Equations (10) to (14) constitute an empirical model of λ_{sat}. Table 3 shows that using θ_{satMOD} (Eqs. 13–14) instead of the θ_{sat} observations has little impact on the λ_{satMOD} scores.

4 Discussion

4.1 Impact of gravels and SOM on q and λ_{sat}

Gravels and SOM are often neglected in soil thermal conductivity models used in LSMs. Moreover, it is often assumed that q is equal to f_{sand}. The Eqs. (10)–(14) empirical model obtained in Sect. 3.2 permits the assessment of the impact of q, f_{gravel} and f_{SOM} on λ_{sat}. Table 3 shows the impact on λ_{satMOD} scores of imposing null values to f_{gravel} and f_{SOM} and of assuming $q = f_{sand}$. The combination of these assumptions is evaluated, also.

It appears that neglecting gravels ($f_{gravel} = 0\,m^3\,m^{-3}$) has a major impact on λ_{sat}: the modelled λ_{sat} is overestimated (with a mean bias of $+0.15\,W\,m^{-1}\,K^{-1}$) and $r^2 = 0.65$, while the full model is virtually unbiased and presents a r^2 value of 0.87. Neglecting SOM also triggers an overestimation of λ_{sat} ($+0.12\,W\,m^{-1}\,K^{-1}$) but has no impact on r^2. On the other hand, although neglecting SOM while accounting for gravels has no impact on r^2, neglecting SOM tends to amplify the detrimental impact of neglecting gravels: $r^2 = 0.51$ and the mean bias is equal to $+0.41\,W\,m^{-1}\,K^{-1}$. Assuming $q = f_{sand}$ tends to trigger an underestimation of λ_{sat} ($-0.22\,W\,m^{-1}\,K^{-1}$), and to compensate for the bias caused by neglecting SOM. Combining Eq. (12) and Eq. (1), it appears that Eq. (12)

SOILD
2, 737–765, 2015

boils down to $q = 1.075 \times f_{sand}$ for θ_{sat} values close to $0.35\,m^3\,m^{-3}$. For higher θ_{sat} values, q tends to be higher than $1.075 \times f_{sand}$. Since θ_{sat} is higher than $0.35\,m^3\,m^{-3}$ at all the sites (Table 1), the $q = f_{sand}$ assumption tends to underestimate q and, subsequently, λ_{sat}.

Table 3 shows that in the configuration representative of most soil thermal conductivity models currently used in LSMs (i.e. neglecting gravels and SOM while assuming $q = f_{sand}$), only 61 % of the λ_{sat} variance is explained by the model ($r^2 = 0.61$), and λ_{sat} is markedly overestimated (the mean bias is equal to $+0.24\,W\,m^{-1}\,K^{-1}$). The impact of this model configuration is illustrated in Fig. 6 (bottom), together with the impact of $q = f_{sand}$ alone.

Finally, the correlation of the retrieved values of q with the volumetric fraction of solids is analysed in Table 4. Using f_{sand} as a predictor of q gives a r^2 value of 0.66, against 0.78 for Eq. (12). Gravels and silt have the largest impact on r^2 ($r^2 = 0.12$ and $r^2 = 0.38$, respectively).

4.2 Null values of q

Null values of q are obtained for four stations: LHS, SVN, NBN, and PRD (Table 2 and Fig. 4). They correspond to the lowest values of the λ_{sat} retrievals, ranging from 0.52 to $1.11\,W\,m^{-1}\,K^{-1}$. Such values can be encountered for organic soils (e.g. Chen et al., 2012) of for volcanic ash soils (Tarnawski et al., 2009) but are surprising for the soil types considered in this study.

On the other hand, it must be noted that Eq. (12) predicts very low values of q for $f_{clay} + f_{silt} + f_{gravel} + f_{SOM}$ close to $0.65\,m^3\,m^{-3}$. All the retrieved null values of q are obtained below this threshold (Fig. 4), from 0.43 to $0.54\,m^3\,m^{-3}$. Therefore, a possible marked underestimation of either f_{clay}, f_{silt}, f_{gravel}, or f_{SOM} may explain these discrepancies, or the overestimation of f_{sand} or θ_{sat}. Uncertainties in these fractions may be caused by (1) the natural heterogeneity of soil properties, (2) the living root biomass, (3) stones that are not accounted for in the gravel fraction. In particular, during the

SOILD
2, 737–765, 2015

installation of the probes, it was obseved that stones are present at the four stations. Stones are not evenly distributed in the soil, and it is not possible to investigate whether the soil area where the temperature probes were inserted contains stones as it must be left unperturbed.

4.3 Applicability of the new λ_{sat} model

The current global soil digital maps provide information about SOM, gravels and bulk density (Nachtergaele et al., 2012). Therefore, using Eq. (1) and Eqs. (6)–(12) at large scale is possible, and porosity can be derived from Eq. (1).

A key component of the λ_{sat} model proposed in this study is the pedotransfer function for quartz (Eq. 12). This equation should be evaluated for other regions. In particular, hematite has to be considered together with quartz for tropical soils. According to Eq. (12), q is close to 0.7 for sandy soils and in such conditions, one should ensure that $q \leq f_{\text{sand}}$.

While the pedotransfer function we get for θ_{sat} (Eq. 13) is valid for the specific sites considered in this study and is used to conduct the sensitivity study of Sect. 4.1, Eq. (13) cannot be used to predict porosity in other regions.

4.4 Prospects for using soil temperature profiles

A key limitation of the data used in this study is that soil temperature observations (T_i) are recorded with a resolution of $\Delta T_i = 0.1\,^{\circ}\text{C}$ only (see Sect. 2.1).

Since T_i is recorded with a resolution of

$$\Delta T_i = \left| \partial \left(T_i^n - T_i^{n-1} \right) \right| = \left| \partial \left(T_{i+1}^n - T_i^n \right) \right| = 0.1\,^{\circ}\text{C}, \tag{15}$$

SOILD

2, 737–765, 2015

the retrieved D_h values are affected by uncertainties and the relative uncertainty of D_h can be estimated as:

$$\left|\frac{\partial D_{hi}}{D_{hi}}\right| = \Delta T_i \times \left\{ \frac{1}{\left|T_i^n - T_i^{n-1}\right|} + \frac{\Delta z_{i+1}^{-1} + \Delta z_i^{-1}}{\left|\gamma_{i+1}^n - \gamma_i^n\right| + \left|\gamma_{i+1}^{n-1} - \gamma_i^{n-1}\right|} \right\}. \tag{16}$$

Therefore, D_h retrievals are more accurate in conditions when soil temperature at $z_i = -0.10$ m changes rapidly and when differences in vertical gradients of soil temperature above and below z_i are more pronounced. In general, this occurs around noon (between 09:00 and 14:00 LST), and at dusk to a lesser extent, between 17:00 and 00:00 LST. In this study, we have imposed the following conditions for using the obtained D_h retrievals:

$$\left|T_i^n - T_i^{n-1}\right| > 0.8\,^\circ\mathrm{C},\ \left|\gamma_{i+1}^n - \gamma_i^n\right| > 30\,\mathrm{K\,m^{-1}},\ \text{and}\ \left|\gamma_{i+1}^{n-1} - \gamma_i^{n-1}\right| > 30\,\mathrm{K\,m^{-1}}. \tag{17}$$

According to Eq. (7), this ensures that

$$\left|\frac{\partial D_{hi}}{D_{hi}}\right| < 18\,\%. \tag{18}$$

It can be noticed that if T_i were recorded with a resolution of 0.03 °C (which corresponds to the typical uncertainty of PT100 probes), D_h could be retrieved with a precision of about 5 % in the conditions of Eq. (17). Alternatively, Eq. (17) conditions could be relaxed in order to get more values of λ estimates for $S_\mathrm{d} > 0.4$ (Sect. 2.6) and increase the number of usable stations. Therefore, one may recommend to revise the current practise of most observation networks consisting in recording soil temperature with a resolution of 0.1 °C only.

SOILD
2, 737–765, 2015

5 Conclusions

An attempt was made to use routine soil temperature and soil moisture observations of a network of automatic weather stations to retrieve instantaneous values of the soil thermal conductivity at a depth of 0.10 m. The data from the SMOSMANIA network, in southern France, are used. First, the thermal diffusivity is derived from consecutive measurements of the soil temperature at a 48-minute interval, at three depths (0.05, 0.10, and 0.20 m). The thermal diffusion equation is solved using an implicit scheme. It is shown that, as the soil temperature records have a resolution of 0.1 °C, the thermal diffusivity can be obtained with a relative error of 18 %. The λ values are then derived from the thermal diffusivity retrievals and from the volumetric heat capacity calculated using measured soil properties. The relationship between the λ estimates and the measured soil moisture at a depth of 0.10 m permits the retrieval of λ_{sat} for 15 stations. A classic λ model is then used to retrieve the quartz volumetric content by reverse modelling. For four stations, low values of λ_{sat} and null values of q are obtained, probably in relation to uncertainties in the gravel and stone fraction. For eleven stations, a pedotransfer function is proposed for quartz, for the considered region in France. A sensitivity study shows that gravels have a major impact on λ_{sat} and that omitting the SOM information tends to enhance this impact. This technique is easy to implement and is based on fully automatic in situ observations associated to a characterisation of soil properties in the lab. Therefore, this study could be extended to other regions and biomes. However, using temperature records with a resolution of 0.1 °C limits the applicability of the method. It is recommended to acquire temperature measurements with a better resolution. More precision in the λ estimates would then permit investigating other processes of heat transfer in the soil such as those related to water transport (Rutten, 2015).

Acknowledgements. We thank our Meteo-France colleagues for their support in collecting and archiving the SMOSMANIA data: Catherine Bienaimé, Marc Bailleul, Laurent Brunier, Anna Chaumont, Jacques Couzinier, Mathieu Créau, Philippe Gillodes, Sandrine Girres, Michel Gou-

SOILD
2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

verneur, Maryvonne Kerdoncuff, Matthieu Lacan, Pierre Lantuejoul, Dominique Paulais, Fabienne Simon, Dominique Simonpietri, Marie-Hélène Théron, Marie Yardin.

References

Abu-Hamdeh, N. H. and Reeder, R. C.: Soil thermal conductivity: effects of density, moisture, salt concentration, and organic matter, Soil Sci. Soc. Am. J., 64, 1285–1290, 2000.

Albergel, C., Rüdiger, C., Pellarin, T., Calvet, J.-C., Fritz, N., Froissard, F., Suquia, D., Petitpa, A., Piguet, B., and Martin, E.: From near-surface to root-zone soil moisture using an exponential filter: an assessment of the method based on in-situ observations and model simulations, Hydrol. Earth Syst. Sci., 12, 1323–1337, doi:10.5194/hess-12-1323-2008, 2008.

Albergel, C., Calvet, J.-C., de Rosnay, P., Balsamo, G., Wagner, W., Hasenauer, S., Naeimi, V., Martin, E., Bazile, E., Bouyssel, F., and Mahfouf, J.-F.: Cross-evaluation of modelled and remotely sensed surface soil moisture with in situ data in southwestern France, Hydrol. Earth Syst. Sci., 14, 2177–2191, doi:10.5194/hess-14-2177-2010, 2010.

Calvet, J.-C., Fritz, N., Froissard, F., Suquia, D., Petitpa,A., and Piguet, B.: In situ soil moisture observations for the CAL/VAL of SMOS: the SMOSMANIA network, International Geoscience and Remote Sensing Symposium, IGARSS, Barcelona, Spain, 23–28 July 2007, 1196–1199, doi:10.1109/IGARSS.2007.4423019, 2007.

Chen, Y. Y., Yang, K., Tang, W., Qin, J., and Zhao, L.: Parameterizing soil organic carbon's impacts on soil porosity and thermal parameters for Eastern Tibet grasslands, Sci. China Earth Sci., 55, 1001–1011, doi:10.1007/s11430-012-4433-0, 2012.

Côté, J. and Konrad, J.-M.: A generalized thermal conductivity model for soils and construction materials, Can. Geotech. J., 42, 443–458, doi:10.1139/T04-106, 2005.

Draper, C., Mahfouf, J.-F., Calvet, J.-C., Martin, E., and Wagner, W.: Assimilation of ASCAT near-surface soil moisture into the SIM hydrological model over France, Hydrol. Earth Syst. Sci., 15, 3829–3841, doi:10.5194/hess-15-3829-2011, 2011.

Johansen, O.: Thermal conductivity of soils. Ph.D. thesis, University of Trondheim, 236 pp., available from Universitetsbiblioteket i Trondheim, Høgskoleringen 1, 7034 Trondheim, Norway, a translation is available at: http://www.dtic.mil/cgi-bin/GetTRDoc?AD=ADA044002 (last access: June 2015), 1975.

Lu, S., Ren, T., Gong, Y., and Horton, R.: An improved model for predicting soil thermal conductivity from water content at room temperature, Soil Sci. Soc. Am. J., 71, 8–14, doi:10.2136/sssaj2006.0041, 2007.

Nachtergaele, F., van Velthuize, H., Verelst, L., Wiberg, D., Batjes, N., Dijkshoorn, K., van Engelen, V., Fischer, G., Jones, A., Montanarella, L., Petri, M., Prieler, S., Teixeira, E., and Shi, X.: Harmonized World Soil Database, Version 1.2, FAO/IIASA/ISRIC/ISS-CAS/JRC, FAO, Rome, Italy and IIASA, Laxenburg, Austria, available at: http://webarchive.iiasa.ac.at/Research/LUC/External-World-soil-database/HWSD_Documentation.pdf (last access: June 2015), 2012.

Parlange, M. B., Cahill, A. T., Nielsen, D. R., Hopmans, J. W., and Wendroth, O.: Review of heat and water movement in field soils, Soil Till. Res., 47, 5–10, 1998.

Parrens, M., Zakharova, E., Lafont, S., Calvet, J.-C., Kerr, Y., Wagner, W., and Wigneron, J.-P.: Comparing soil moisture retrievals from SMOS and ASCAT over France, Hydrol. Earth Syst. Sci., 16, 423–440, doi:10.5194/hess-16-423-2012, 2012.

Peters-Lidard, C.D., Blackburn, E., Liang, X., and Wood, E.F.: The effect of soil thermal conductivity parameterization on surface energy fluxes and temperatures, J. Atmos. Sci., 55, 1209–1224, 1998.

Rutten, M. M.: Moisture in the topsoil: From large-scale observations to small-scale process understanding, PhD Thesis, Delft university of Technology, doi:10.4233/uuid:89e13a16-b456-4692-92f0-7a40ada82451, available at: http://repository.tudelft.nl/view/ir/uuid:89e13a16-b456-4692-92f0-7a40ada82451/ (last access: June 2015), 2015.

Schelde, K., Thomsen, A., Heidmann, T., Schjonning, P., and Jansson, P.-E.: Diurnal fluctuations of water and heat flows in a bare soil, Water Resour. Res., 34, 2919–2929, 1998.

Schönenberger, J., Momose, T., Wagner, B., Leong, W. H., and Tarnawski, V. R.: Canadian field soils I. Mineral composition by XRD/XRF measurements, Int. J. Thermophys., 33, 342–362, doi:10.1007/s10765-011-1142-4, 2012.

Tarnawski, V. R., McCombie, M. L., Leong, W. H., Wagner, B., Momose, T., and Schönenberger J.: Canadian field soils II. Modeling of quartz occurrence, Int. J. Thermophys., 33, 843–863, doi:10.1007/s10765-012-1184-2, 2012.

Tarnawski, V. R., Momose, T., and Leong, W. H.: Assessing the impact of quartz content on the prediction of soil thermal conductivity, Géotechnique, 59, 4, 331–338, doi:10.1680/geot.2009.59.4.331, 2009.

SOILD

2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

Yang, K., Koike, T., Ye, B., and Bastidas, L.: Inverse analysis of the role of soil vertical heterogeneity in controlling surface soil state and energy partition, J. Geophys. Res., 110, D08101, doi:10.1029/2004JD005500, 2005.

SOILD
2, 737–765, 2015

Table 1. Soil characteristics at 10 cm for the 21 stations of the SMOSMANIA network. Porosity values are derived from Eq. (1). Solid fraction values higher than 0.3 are in bold. The stations are listed from West to East (from top to bottom). ρ_d, θ_{sat}, f, and m, stand for soil bulk density, porosity, volumetric fractions, and gravimetric fractions, respectively.

Station	ρ_d ($kg\,m^{-3}$)	θ_{sat} ($m^3\,m^{-3}$)	f_{sand} ($m^3\,m^{-3}$)	f_{clay} ($m^3\,m^{-3}$)	f_{silt} ($m^3\,m^{-3}$)	f_{gravel} ($m^3\,m^{-3}$)	f_{SOM} ($m^3\,m^{-3}$)	m_{sand} ($kg\,kg^{-1}$)	m_{clay} ($kg\,kg^{-1}$)	m_{silt} ($kg\,kg^{-1}$)	m_{gravel} ($kg\,kg^{-1}$)	m_{SOM} ($kg\,kg^{-1}$)
SBR	1680	0.352	**0.576**	0.026	0.013	0.002	0.032	**0.911**	0.041	0.020	0.003	0.024
URG	1365	0.474	0.076	0.078	**0.341**	0.005	0.025	0.149	0.153	**0.665**	0.009	0.024
CRD	1435	0.438	**0.457**	0.027	0.033	0.000	0.045	**0.848**	0.051	0.060	0.000	0.041
PRG	1476	0.431	0.051	0.138	0.138	0.214	0.028	0.092	0.250	0.248	**0.385**	0.025
CDM	1522	0.413	0.073	0.241	0.231	0.012	0.030	0.128	**0.422**	**0.404**	0.020	0.026
LHS	1500	0.416	0.102	0.202	0.189	0.051	0.039	0.181	**0.359**	**0.335**	0.091	0.034
SVN	1453	0.445	0.127	0.073	0.176	0.162	0.017	0.233	0.133	**0.322**	0.296	0.015
MNT	1444	0.447	0.135	0.066	0.230	0.102	0.020	0.248	0.121	**0.424**	0.188	0.018
SFL	1533	0.413	0.127	0.071	0.118	0.250	0.021	0.221	0.123	0.205	**0.434**	0.018
MTM	1540	0.405	0.110	0.081	0.076	0.297	0.032	0.189	0.140	0.131	**0.512**	0.027
LZC	1498	0.429	0.129	0.066	0.068	0.292	0.015	0.229	0.117	0.121	**0.519**	0.013
NBN	1545	0.401	0.063	0.135	0.075	0.290	0.035	0.109	0.232	0.130	**0.499**	0.030
PZN	1311	0.495	0.222	0.074	0.131	0.054	0.023	**0.450**	0.151	0.266	0.111	0.023
PRD	1317	0.494	0.038	0.052	0.069	**0.326**	0.021	0.076	0.105	0.139	**0.659**	0.021
LGC	1496	0.428	0.253	0.044	0.042	0.214	0.019	**0.451**	0.078	0.074	**0.380**	0.017
MZN	1104	0.560	0.212	0.037	0.045	0.097	0.049	**0.510**	0.089	0.109	0.234	0.057
VLV	1274	0.506	0.294	0.054	0.086	0.031	0.029	**0.614**	0.112	0.179	0.064	0.030
BRN	1630	0.379	0.105	0.009	0.016	**0.474**	0.016	0.171	0.015	0.027	**0.774**	0.013
MJN	1276	0.506	0.064	0.029	0.056	**0.317**	0.028	0.133	0.060	0.118	**0.661**	0.029
BRZ	1280	0.508	0.097	0.074	0.109	0.190	0.020	0.202	0.154	0.228	**0.396**	0.021
CBR	1310	0.501	0.120	0.057	0.068	0.241	0.013	0.243	0.116	0.139	**0.489**	0.013

SOILD
2, 737–765, 2015

SOILD
2, 737–765, 2015

Table 2. λ_{sat} and q retrievals using the λ model (Eqs. 7–9 and Eq. 10, respectively) for degree of saturation values higher than 0.4, together with the minimized RMSD between the simulated and observed λ values, and the number of used λ observations (n). The four stations with $q = 0$ are in bold.

Station	Station full name	λ_{sat} ($W\,m^{-1}\,K^{-1}$)	q ($m^3\,m^{-3}$)	RMSD ($W\,m^{-1}\,K^{-1}$)	n
SBR	SABRES	2.79	0.61	0.233	118
URG	URGONS	1.28	0.13	0.102	62
CRD	CREON-D'ARMAGNAC	–	–	–	0
PRG	PEYRUSSE-GRANDE	1.59	0.26	0.105	94
CDM	CONDOM	1.36	0.13	0.263	132
LHS	**LAHAS**	**1.03**	**0.00**	**0.405**	**116**
SVN	**SAVENES**	**1.11**	**0.00**	**0.276**	**997**
MNT	MONTAUT	1.45	0.38	0.103	207
SFL	SAINT-FELIX-DE-LAURAGAIS	1.40	0.14	0.183	1078
MTM	MOUTHOUMET	1.45	0.18	0.137	1939
LZC	LEZIGNAN-CORBIERES	1.49	0.19	0.166	557
NBN	**NARBONNE**	**0.52**	**0.00**	**0.101**	**100**
PZN	PEZENAS	–	–	–	0
PRD	**PRADES-LE-LEZ**	**0.70**	**0.00**	**0.165**	**490**
LGC	LA-GRAND-COMBE	1.80	0.34	0.368	305
MZN	MAZAN-L'ABBAYE	–	–	–	0
VLV	VILLEVIEILLE	–	–	–	0
BRN	BARNAS	1.35	0.07	0.335	469
MJN	MEJANNES-LE-CLAP	–	–	–	0
BRZ	BERZEME	–	–	–	0
CBR	CABRIERES-D'AVIGNON	1.72	0.36	0.241	85

Table 3. Ability of the Eqs. (10)–(13) empirical model to estimate λ_{sat} values at the 11 stations presenting non-null q values and impact of changes in the model, in order of decreasing r^2 score (from top to bottom). Results for all 15 stations (including null q values) are between brackets.

Model configuration	r^2	RMSD ($W\,m^{-1}\,K^{-1}$)	Mean bias ($W\,m^{-1}\,K^{-1}$)
Full model using θ_{sat} observations	0.87 (0.03)	0.15 (1.08)	−0.01 (+0.62)
Full model using θ_{satMOD} (Eqs. 13–14)	0.88 (0.01)	0.14 (1.02)	−0.03 (+0.62)
same with: $f_{SOM} = 0$	0.87 (0.00)	0.23 (1.18)	+0.12 (+0.78)
same with: $f_{SOM} = 0$ and $q = f_{sand}$	0.86 (0.02)	0.22 (1.06)	+0.00 (+0.64)
same with: $q = f_{sand}$	0.86 (0.12)	0.26 (0.82)	−0.22 (+0.37)
same with: $f_{gravel} = 0$ and $q = f_{sand}$	0.77 (0.17)	0.25 (0.81)	−0.15 (+0.39)
same with: $f_{gravel} = 0$	0.65 (0.07)	0.29 (1.06)	+0.15 (+0.73)
same with: $f_{gravel} = 0$, $f_{SOM} = 0$ and $q = f_{sand}$	0.61 (0.04)	0.42 (1.17)	+0.24 (+0.83)
same with: $f_{gravel} = 0$ and $f_{SOM} = 0$	0.51 (0.01)	0.56 (1.34)	+0.41 (+1.01)

SOILD
2, 737–765, 2015

Table 4. Coefficient of determination and RMSD of the regression equation of q vs. diverse volumetric fractions of soil solids, for 11 SMOSMANIA stations. First line corresponds to the pedotransfer function for quartz proposed in this study (Eq. 12).

Predictor of q	r^2	RMSD ($m^3\,m^{-3}$)
$f_{clay} + f_{silt} + f_{gravel} + f_{SOM}$	0.78	0.07
$f_{clay} + f_{silt} + f_{gravel}$	0.78	0.07
$f_{silt} + f_{gravel} + f_{SOM}$	0.67	0.09
f_{sand}	0.66	0.09
$f_{clay} + f_{gravel} + f_{SOM}$	0.38	0.12
$f_{clay} + f_{silt} + f_{SOM}$	0.12	0.14
$f_{sand} + f_{gravel}$	0.11	0.14

SOILD
2, 737–765, 2015

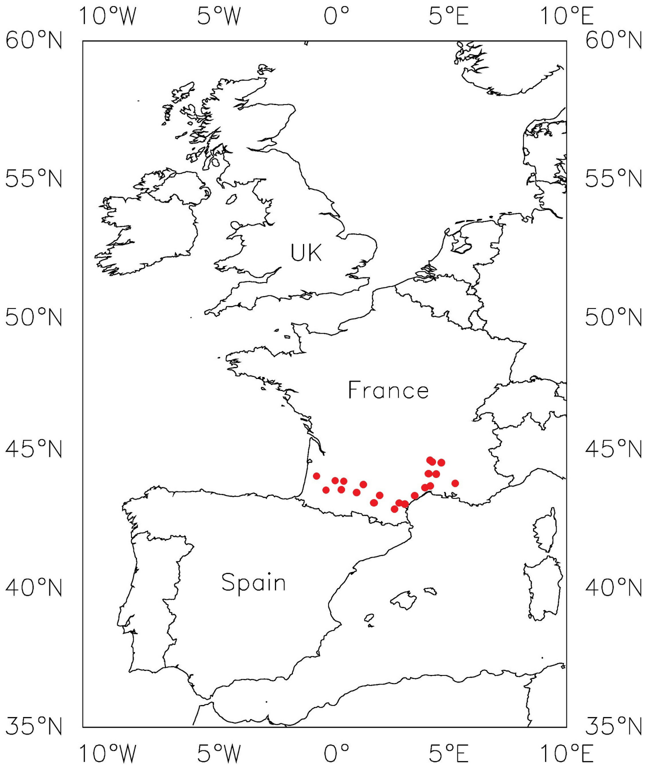


Figure 1. Location of the 21 SMOSMANIA stations in southern France.

SOILD
2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

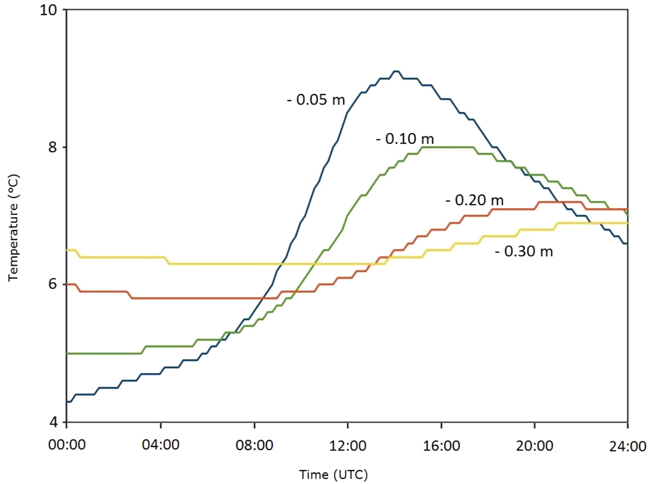


Figure 2. Soil characteristics of the 21 SMOSMANIA stations at a depth of 10 cm: mineral fine earth gravimetric fractions of clay, silt and sand. For a given station, the red mark covers the fraction values measured at 0.05, 0.10 and 0.20 m. Full station names are given in Table 2. The dashed blue lines correspond to the USDA textural soil classification.

SOILD
2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

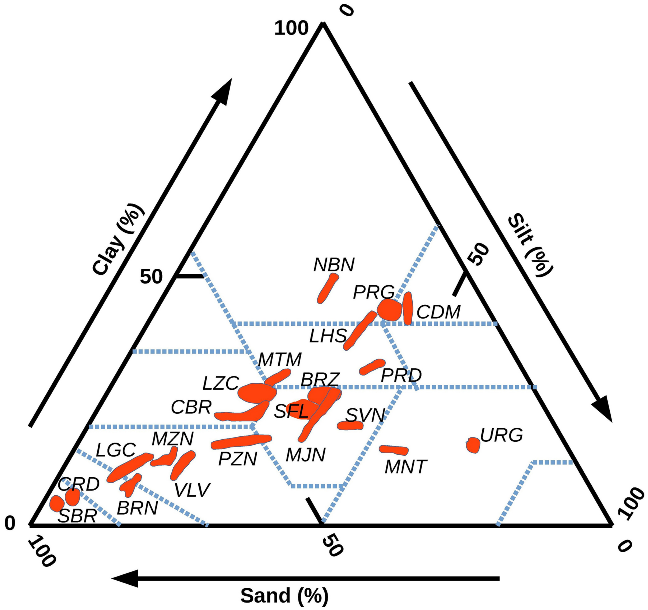


Figure 3. Soil temperature measured at the Saint-Félix-de-Lauragais (SFL) station on 23 February 2015, at depths of 0.05, 0.10, 0.20, and 0.30 m. Levelling is due to the low resolution of the temperature records (0.1 °C).

SOILD

2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

Title Page
Abstract | Introduction
Conclusions | References
Tables | Figures
◀I | ▶I
◀ | ▶
Back | Close
Full Screen / Esc
Printer-friendly Version
Interactive Discussion

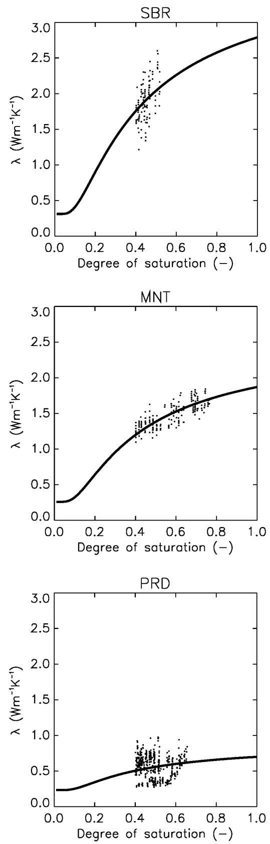


Figure 4. Retrieved λ values vs. the observed degree of saturation of the soil, at a depth of 0.10 m, for (from top to bottom) Sabres (SBR), Montaut (MNT), and Prades-le-Lez (PRD).

SOILD
2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.

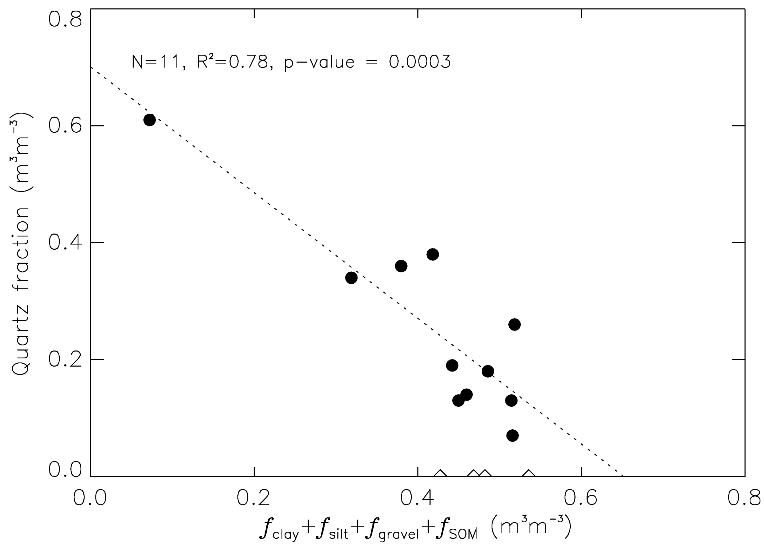


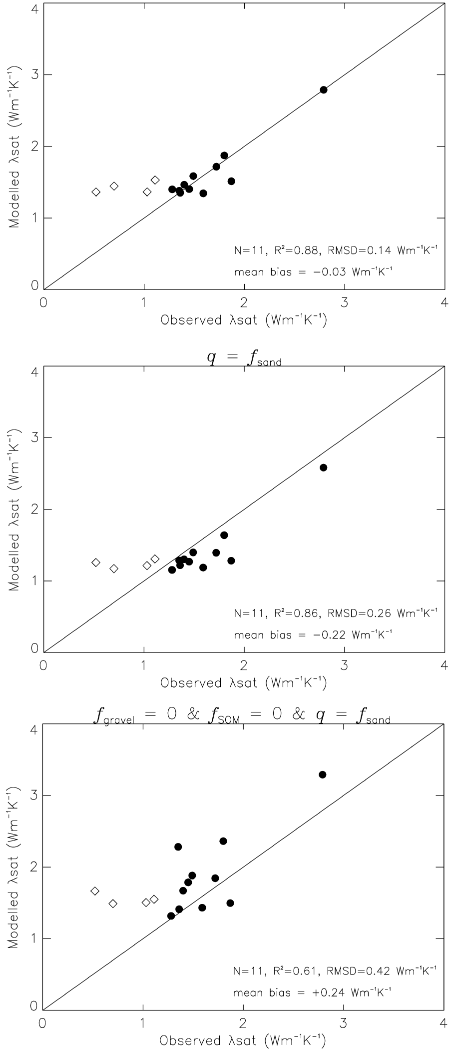
Figure 5. Non-null and null q retrievals (dark dots and opened diamonds, respectively) vs. the sum of clay, silt, gravel and SOM volumetric fractions (empirical Eq. 12).

SOILD

2, 737–765, 2015

Impact of gravels and organic matter on the thermal properties of grassland soils

J.-C. Calvet et al.



SOILD
2, 737–765, 2015

Figure 6. λ_{satMOD} vs. λ_{sat} retrievals for non-null and null q retrievals (dark dots and opened diamonds, respectively): (top) full model using θ_{satMOD} (Eqs. 13–14), (middle) replacing Eq. (12) by $q = f_{sand}$, (bottom) replacing Eq. (12) by $q = f_{sand}$ and assuming $f_{gravel} = 0$ and $f_{SOM} = 0$ (Table 3).

SOILD
2, 737–765, 2015