FACTORS AFFECTING SOIL TEMPERATURE AS LIMITS OF SPATIAL INTERPRETATION AND SIMULATION OF SOIL TEMPERATURE

Michal Lehnert [1]

[1] Department of Physical Geography and Geoecology, University of Ostrava, Chittussiho 10 Ostrava, Czech Republic, michal.lehnert@gmail.com

Abstract

Soil temperature influences the course of many important processes across the landscape. Nevertheless, the studies performed of it so far have made only small progress in the field of soil physics, as well as in climate studies. This paper systematically summarises the hitherto vaguely arranged pieces of knowledge on the factors affecting the soil temperature and related approaches to spatial interpretation and simulation of soil temperature. Subsequently, two main problems limiting the development of the scientific study of soil temperature have been identified: 1) methodological inconsistencies in the study of soil temperature and factors affecting soil temperature and 2) the low density of points with a representative measurement of soil temperature. Because of the high spatial heterogeneity and the complexity of the relationships of the factors influencing soil temperature, any further studies of soil temperatures must develop in a methodically coherent manner. To achieve this, it seems appropriate to use the concept of elementary process-oriented geo-systems.

Key words: soil temperature, field heterogeneity, soil modelling.

INTRODUCTION

Soil temperature has a major influence on the course of many important processes across the landscape and it is an important factor in pedogenesis (Jenny 1941). It controls intraspheric metabolic processes such as e.g. the weathering of minerals (Wu and Nofziger 1999), the decomposition of organic compounds (Davidson and Janssens 2006) or the mineralisation of nitrogen (Gonçalves and Carlyle 1994) and degradation and transport of pollutants (Wu and Nofziger 1999). It significantly affects the interspheric processes of gas exchange between the atmosphere and soil (Lloyd and Taylor 1994) or plant growth and germination (Šulgin 1967; Bedrna 1984; Probert 2000). Despite some partial criticism (e.g. Smith 1973), soil temperature regimes remain components of important soil taxonomic systems such as the Soil Survey Staff (most recently USDA 2014) and, to a lesser extent, WRB (most recently FAO 2014) and are also an important characteristic for assessing the quality and fertility of the soil, e.g. BPEJ (Maštát 2002).

Nosek (1972) mentioned that the study of soil temperature does not receive sufficient attention. Bedrna (1989) stated that the importance of soil temperature has not been fully appreciated. Možný (1991) notes the lack of climatological stations measuring soil temperature for assessing soil temperature regimes. Buchan (2001) aptly notes that although soil temperature affects virtually all the terrestrial biosphere, it has still been studied very little.

Despite the significant progress of computing capabilities in recent years, in the interpretation and prediction of the spatial variability of soil temperatures we still find a high degree of scientific uncertainty. Therefore, the goal of this paper is to provide a systematic summary of key findings on factors affecting soil temperature and related approaches to the spatial simulation and prediction of soil temperature. Finally, it assesses the current research into soil temperatures and outlines its problems and prospects.

FACTORS AFFECTING SOIL TEMPERATURE

Hanks (1992) divides the factors affecting thermal soil regimes into those that affect the heat available at the soil surface and those affecting the transmission of heat in the soil. Green et al. (1984) suggest another classification based on distinguishing global and local factors that influence soil temperature. It is mainly British and Northern European agrometeorologists (e.g. Green and Harding 1979, 1980; Green et al. 1984; Oliver et al. 1987) who supported the approach based on the observation of Bouyoucos (1913) that the physical properties of the soil have less effect on soil temperature than expected. On the other hand, they emphasise the influence of weather conditions and the properties of the active surface, especially vegetation. Generally, however, the Euro-American space – as the crucial domain of research into the soil temperature regime – has become dominated by the approaches of classical soil physics based on works by members of the "Wageningen School" (Philip and de Vries 1957; de Vries 1963; van Wijk and de Vries 1963; van Wijk 1963).

These studies interpret temperature in the context of the hydrothermal regimes of the soil. These analytical approaches to the study of the thermal properties of soils were also popular with authors from the former Eastern bloc (e.g. Chudnovsky 1976). The current approaches to the study of soil temperature look at this issue more comprehensively with regard to all the factors that affect soil temperature (e.g. Zhang 2003). Considering the complex system of feedback, any classification of the factors affecting soil temperature has a rather conceptual meaning. In this study, the factors affecting soil temperature are divided into those relating to soil environment properties, land relief and land cover factors and the factors relating to the state of the atmosphere. The impact of human activity affects all the abovementioned categories (Table 1). A description of the factors which condition the purely anthropogenic soil thermal regimes as suggested by Dimo (1972), i.e. irrigated soils, heated soils, soils overlaid with artificial materials, and buried soils (including soils in roofed spaces), is beyond the scope of this study. It should also be noted that as the factors that affect the mesospheric thermic regime of the soil always simultaneously influence the conditioned interspheric soil temperature (thermal) regime, the survey also includes works which examine the issue from the perspective of thermic regimes.

Properties of the soil environment

Considering the primary focus of most of the relevant work on the thermal properties, the impacts of the physical properties of the soil on soil temperature can be advantageously summarised within the context of thermal conductivity, heat capacity and thermal diffusivity. All these physical features inherently affect soil temperature (see above).

The thermal conductivity of soil is mainly influenced by the volume fraction of solid, liquid and gaseous substances. Therefore, for the determination of the thermal conductivity of soil, its porosity or bulk density are essential, as is the actual moisture content (e.g. de Vries 1952; Al and Nakshabandi Köhnke 1965; Ochsner et al. 2001; Lu et al. 2007).

The thermal conductivity is also influenced by the actual composition of the solid fraction – the type of minerals, size of the soil constituents and the amount of organic substances. Kersten (1949) was one of the first to compare the thermal conductivity of individual soil constituents. He pointed out, for example, the higher thermal conductivity of silica than of plagioclase feldspar or pyroxene, and in particular, the significantly higher thermal conductivity of ice than water (which is an important factor in the process of soil freezing). De Vries (1963)

Table 1 Systematic summary of the key findings of factors influencing the soil temperature.

Interactios	Factors	Anthropogenic influence
	SOLAR RADIATION	
	State of the atmosphere *Pressure gradients and wind:* Rettig (1956); Hanks et al. (1967); Cellier et al. (1996) *Transmittance of atmosphere:* Budyko (1956); Lacis and Hansen (1974)	
	Land relief and surface properties *Geometric and topographic characteristics of the relief:* Grunow (1952); Krcho (1965); Kondratyev (1969); Harding (1979,1980); Bedrna (1980) *Land cover character:* List (1959); Šulgin (1972); Skartveit (1976); Idso et al. (1975); Kondratyev et al. (1982); Bedrna, Gašparovič 1985; Oliver et al. (1987); Matthias et al. (2000); Decker et al. (2003)	
	Soil properties *Water content and state:* Kersten (1952); De Vries (1952); Al Nakshabandi and Kohnke (1965); Yadav and Saxena (1973); Sepaskhah and Boersma (1979); Ghumam and Lal (1985); Abu-Hamdeh (2003); Balland and Arp (2005); Lu et al. (2007) *Porosity + bulk density + grain size:* de Vries (1952); Al Nakshabandi and Kohnke (1965); Tavman (1996); Abu-Hamdeh (2003); Balland and Arp (2005) *Chemical composition:* Kersten (1949); De Vries (1963); Sepaskhah and Boersma (1979); Noborio and McInnes (1993); Abu-Hamdeh and Reeder (2000) *Soil temperature:* de Vries (1963); Sepaskhah and Boersma (1979) *Water and water vapour convection:* Ramdas and Dravid (1934); Brooks and Rhoades (1953); Philip and de Vries (1957); Johansen (1977); Thomas and He (1995)	
	SOIL TEMPERATURE	

compiled a table of the thermal conductivity of the basic soil-forming components, which shows that the thermal conductivity of silica is almost three times higher than the thermal conductivity of the clay minerals and nearly thirty times higher than the thermal conductivity of the organic substances. A decrease in thermal conductivity in connection with increasing amounts of organic substances is also confirmed by Abu-Hamdeh and Reeder (2000).

Tavman (1996) presented an exact theory of increasing thermal conductivity according to the increasing size of soil constituents. Abu-Hamdeh and Reeder (2000), on the basis of their results and as a summary of the earlier results of other authors, confirmed that with the same water content sandy soil always has a higher thermal conductivity than sandy loam, clay loam or clay. With the growth of the bulk density and soil moisture, the differences in thermal conductivity between sandy soils and other soils increase. This is related to the size of the contact surfaces at the points of contact of conductive elements. Because of the larger pore size of the contact surface at higher humidity, the contact surfaces extend most in sandy soils. The thermal conductivity begins to increase with the moisture content only if the water films around the soil constituents are connected. Therefore, at low soil moisture, an increase in the thermal conductivity of soil and of the soil moisture is insignificant, but then the thermal conductivity increases until the pores are filled with water (Sepaskhah and Boersma 1979). According to Tarnawski and Leong (2000), a significant increase in the thermal conductivity begins from a moisture content of 0.02-0.26 cm^3/cm^3. These threshold values of the initialisation of an increase in the thermal conductivity vary, depending on the grain size of the soil constituents and the bulk density of the soil constituents, and the thermal conductivity increases if the size of the soil constituents gets smaller. Ghuman and Lal (1985) also found lower thermal conductivity in soils containing gravel than in gravel-free soils. From these findings it is obvious that in dry soils, clay soils and soils with a humus horizon the surface of the soils is overheated while the lower horizons are less warmed up. This is evidenced by the conclusions of Hora (2011) based on the measurements of soil temperature in soils with different grain compositions near Brno (Czech Republic).

Another factor affecting the thermal conductivity of soil is the salt content in the soil solution. With an increase in the salt content in the soil solution the soil conductivity decreases. The salt content reduces the thermal conductivity most in sandy soils and least in clay soils (Noborio and McInnes 1993).

According to de Vries (1963), the thermal conductivity of the soil may also be influenced by soil temperature. The thermal conductivity increases with temperatures up to 60°C. However, Balland and Arp (2005) point out the fact that changes in the thermal conductivity at temperatures of −30 to 30°C are insignificant for most purposes. Nevertheless, soil temperature has an indirect but major impact on the thermal conductivity of the soil if the soil water freezes (Kersten 1952). In addition, in the case of frequent freezing of the soil, the consequent changes in the soil structure are reflected in the thermal properties of the soil (Chamberlain et al. 1979).

From the theory of the heat transport in the soil, it is clear that another key physical property of soil affecting the soil temperature regime is the volumetric heat capacity. Therefore, Kersten (1949) presents an overview of the values of specific heat for selected types of soil, while a summary of the values of the specific heat of minerals and humus occurring in soil is presented by de Vries (1963). Yadav and Saxena (1973) found that, at various water contents and bulk densities, clay soils have a higher heat capacity than sandy soils. Abu-Hamdeh (2003) proved that the heat capacity of the majority of soils increases almost linearly with the water content. The heat capacity of soils therefore depends on the moisture content and the mineral and organic composition of the soil. Consequently, soils with lower water content (sand soil or soil with a thick humus horizon) have a lower daily temperature amplitude than soils with a high water content (clay soil). The growth of the daily amplitude of soil temperature with a share of the sand fraction in fine earth is also indicated by the analysis of soil temperatures carried out by Lehnert (2013) in the surroundings of Olomouc (Czech Republic).

With regard to the temperature regime of soil, it is practical to monitor the thermal diffusivity. The thermal diffusivity achieves its highest intensity in fields with relatively high moisture content, from 0.25-0.30 cm^3/cm^3 (Ghumam and Lal 1985; Arkhangel'skaya and Umarova 2008). Since the thermal diffusivity is defined as the quotient of thermal conductivity and the product of the specific heat capacity and bulk density, a small increase in thermal conductivity is not able to compensate for the almost linear increase in heat capacity, which results in a decrease in the thermal diffusivity from the above-mentioned value. The function expressing the dependence of the thermal diffusivity on the water content in the soil is therefore a combination of an S-curve and a parabola (Arkhangel'skaya and Umarova 2008).

From the above, it is clear that for the interpretation of soil temperature regimes, we cannot analyse the thermal regimes separately from the hydric regimes. The basic mechanisms of the interconnected hydrothermal regimes have been described by Philip and de Vries (1957). Johansen (1977) presents an exact theory of the assessment of the effective thermal conductivity. He emphasises the complex relationships between the values of conduction-based heat transport in soil and the natural/forced convection of water, temperature gradients and the diffusion of water vapour and radiation. The coupled heat and moisture transfer in soil is described in detail by Thomas and He (1995).

Human activities have a considerable influence on both the physical and chemical properties and the thermal properties of the soil and the soil temperature itself. The anthropogenic impacts on the soil are particularly important in urban and agricultural landscapes. Van Duin (1956) synthesised the seemingly contradictory results of several authors and found that tillage had only a small effect on the annual variation in soil temperature. At the same time, however, he described a significant increase in the daily amplitude of soil temperature in arable land. Standard line ploughing indicates slightly higher daily amplitude of soil temperature than chisel ploughing or zero tillage (Johnson and Lowery 1985; Licht snd Al-Kaisi 2005). In the long term, ploughed soils mostly heat up more slowly than soils that are not ploughed (Bedrna 1989). Soil temperature is also significantly influenced by an increase in its thermal conductivity as a result of the pedocompaction of the soil caused by the frequent passage of agricultural machinery. For example, Šulgin (1972), by rolling the soil, increased its temperature at a depth of 3 cm by 1 to 2°C.

Land relief and land cover

The energy balance of the soil depends on the surface of the soil. In the case of a bare soil it consists of the sensible heat into the ground (mainly influenced by the above-described thermal properties of the soil), latent heat (determined mainly by the soil moisture and atmospheric factors listed below) and radiation balance.

The key factors that affect soil temperature through the radiative balance of the soil surface are the inclination, orientation and nature of the surface and its shading by the surrounding relief. The principles of the quantification of the radiation incident on inclined surfaces are comprehensively described by Kondratyev (1969). Another significant work in this context is Krch (1965), who took into account the shading of the surface by the surrounding relief. In this context, we cannot ignore the reflected radiation (Ineichen et al. 1990). Geiger et al. (2003) synthesised the results of several authors and emphasised the often neglected influence of scattered shortwave radiation, which, in appropriate weather conditions (especially various types of cloud formation), may significantly reduce the spatial differences in the values of the incident global radiation caused by different orientation and the gradient of the slope. This fact was first pointed out by Grunow (1952).

Another important factor influencing the active surface energy balance is the albedo. The soil albedo ranges from 0.08 for dark and moist soils to 0.3 for light and dry soils (List 1959). The decrease in the albedo with an increase in the surface soil moisture is described by Idso et al. (1975). Cellier et al. (1996) found that after rain the soil albedo decreases by about 0.1 (depending on the soil type). Kondratyev et al. (1982) determined the albedo of individual soil types and Matthias et al. (2000) described the changes in the soil albedo in dependence on the surface roughness. Bedrna et al. (1989) concluded that dark, wet and bumpy soil absorbs more radiation than light dry flat soil.

In a real landscape, a large part of the soil surface is covered by vegetation. The values of lawn albedo, depending on the density, vegetation stage and composition of the vegetation, range from 0.18 to 0.26 (Gates and Hanks 1967; Kondratyev 1982). The albedo of field crops in temperate conditions ranges from 0.13 to 0.39 for beet (depending on the growing stage), over 0.16 to 0.17 for maize and 0.21 to 0.22 for barley (Gates and Hanks 1967). For coniferous forests, Betts and Ball (1997) give a very low albedo value of 0.083. The albedo of oak forests varies from 0.14 to 0.18, depending on the season or the vegetative stage (DeWalle and McGuire 1973). Unlike the albedo, the spatial variability of reflectance (long-wave albedo) of natural surfaces is lower than 0.05 (Gayevsky 1951).

For all growing crops, it is necessary to take into account not only the reflection but also the absorption and transformation of radiation at the plant surface. The radiation which is transmitted through the vegetation to the soil surface depends on the density of the vegetation cover and the ratio of its ability to absorb, transform and reflect the incident radiation. Russel et al. (1990) emphasise that the amount of radiation on the soil surface under the vegetation is also influenced by the angle of the incident radiation. Rodskjer et al. (1989) describe a gradual decrease in the intensity of the heat transfer into the ground with an increase in leaf area index during the growing season under barley. Oliver et al. (1987) calculated the ratio of the heat transfer to the ground and the net radiation over the active surface and confirmed that this ratio varies with the stages of plant vegetation and soil conditions. The transfer of heat into the soil can reach up to 50% of the net radiation. If vegetation cover is present (grass, arable land), it reaches only about 10% of the net radiation (Monteith 1958, Choudhury et al. 1987). Hutchison and Baldocchi (1989) found that the ratio of the net radiation on the surface of the soil and the net radiation at the upper level of vegetation is less than 5% for oak forests and reaches almost 30% in the pine forests of the temperate zone. Green et al. (1984) note that if the vegetation is sufficiently dense and forms a soil-vegetation continuum, the surface of the vegetation has higher temperature amplitudes than the actual surface of the soil and the soil under the dense vegetation is therefore cooler in summer and warmer in winter. Therefore, vegetation acts as a thermal insulator and significantly affects the soil temperature regime, as evidenced by Sándor and Fodor (2012), who observed the decrease in the daily amplitude of soil temperature with the increasing leaf area index in a stand of corn.

Another significant influence on the radiation and energy balance of the surface is that of withered vegetation. Horton et al. (1996) concluded that crop residues change the albedo and emissivity of the surface, weaken the value of the net radiation on the very surface of the soil, inhibit evaporation and increase soil moisture. Consequently, leaf litter reduces the temperature extremes on the soil surface (Manrique 1988; Shinners et al. 1994). Therefore, straw mulching also reduces the average soil temperature (Šulgin 1972; Dahiya et al. 2007).

In addition to the described direct impact of vegetation on the radiation and energy balance of the surface, the vegetation also affects soil temperature indirectly. Minor losses of latent heat with an increase in the leaf area index as a result of lower wind speed and lower turbulence were described by Baldocchi et al. (2000). Therefore, the specific local climate in stands of trees is characterised by its own moisture regime (in detail e.g. Jarvis 1985; Středová et al. 2011).

The relationships between surface properties and vegetation and soil temperature have been described by a wide range of authors. Oliver et al. (1987) found daily amplitudes of a bare soil surface on sandy soils near Thenford (Great Britain) in August of 13.4°C, while on a regularly mown lawn the figure was 7.2°C and on the soil surface in a pine forest only 2.6°C, although the average temperatures of all the surfaces differed only a little. Bedrna and Gašparovič (1985) found that in the Danube basin (Slovakia) the temperature of the soil under the bare surface is, on average, lower by 0.1 to 1.4° eT than under mown grassy vegetation. Možný (1991), on the basis of long-term measurements of soil temperatures at a depth of 20 cm below the surface in Doksany (Czech Republic), found higher temperatures under a bare surface than under a regularly mown lawn as follows: in April by 0.4°C, in May by 1.0°C, in June and July by 0.8°C and in August by 0.7°C. At a depth of 20 cm in Ultuna (Sweden), Rodskjer et al. (1989) found average soil temperatures 1.4°C higher under a wheat crop than under bare soil. Toogood (1979) measured higher soil temperatures under barley than under grassy vegetation near Edmonton (Canada) in the period from late April to late June. In the other months, the soil under grass vegetation was warmer (the harvest had no significant influence on this relationship). In winter, around Edmonton (Canada), Toogood et al. (1979) measured the average soil temperature at a depth of 20 cm below the surface as being 3.0°C higher beneath shrubby surfaces than the temperature of the soil under grass vegetation. Jiménez et al. (2007) describe dramatic differences in the thermal regime (increase in amplitude) on the Canary Islands (Spain) when comparing the original cloud forest and the tree-heath woodland and adjacent areas where these forests were replaced by pine forest, herbaceous plants and cropping and herbaceous plants. Coufal et al. (1993) reported that even under the well-kept lawns at meteorological stations, there are not the same conditions as, in particular, in the surface soil layer (up to 5 cm); he found differences in soil temperature according to the variations in the representation of plant species and the current height and density of the lawn.

Snow cover functions as a heat insulator in the same manner as vegetation. The daily amplitude of soil temperature in the presence of a sufficiently deep snow cover is almost zero. Skartveit (1976) notes that in the presence of snow cover, soil temperature is maintained at a temperature close to 0°C, while the air temperature may be much lower. Decker et al. (2003) describes a higher soil temperature under snow cover during cold winters and in areas where the snow cover was removed during warm winters. The impacts of different levels of snow cover on soil temperature in the Canadian boreal zone were observed by Zhang et al. (2008). In Central Europe, similar effects were observed by Pokladníková et al. (2008).

Following numerous measurements in Western Europe, Green and Harding (1979, 1980) described a drop in soil temperature with altitude. They found that the drop in soil temperature with altitude is generally smoother than the drop in the air temperature. At the same time, however, they pointed out that the changes in soil temperature with altitude vary in the regions that were surveyed. In Slovakia, Bedrna (1980) found that soil temperature at a depth of 20 cm drops by 0.4°C on average per 100 metres of altitude during the growing period. At the same time, he pointed out that this change is closely related to the distance from the mountains and the division of the relief rather than to the actual altitude. In Georgia, Elisbarashvili et al. (2007, 2010) identified a different drop in soil temperatures with altitude in various soil types. They explain the drop in soil temperature with altitude by higher wind speeds in the context of increased evaporation, higher amounts of precipitation, greater cloudiness and the more frequent occurrence of local temperature inversions. The influence of atmospheric temperature inversions on soil temperature regimes around Olomouc (Czech Republic) was also observed by Lehnert (2013). The influence of meteorological or climatic factors on soil temperature is always linked with the influences of relief and therefore modified by local climatic effects (Vysoudil 2009).

State of the atmosphere

The value of the surface net radiation, provided that the distance between the Earth and the sun is constant, the solar activity invariable and the surface horizontal, primarily depends on the height of the sun above the horizon and the actual transmittance of the atmosphere (see e.g. Budyko 1956; Lacis and Hansen 1974). Although radiation is the predominant soil heat source, we cannot find a clear dependence between its intensity and soil temperature. Valuš (1974) claims that only a small part of the variability of the soil average temperature regime at a depth of 20 cm can be directly dependent on the intensity of the radiation.

Other impacts of meteorological (climatic) factors on soil temperature remain hidden in the mechanisms of heat propagation in the soil. One example is the relationship between soil temperature and air temperature. The heat exchange between the atmosphere and the soil by means of conduction is minimal. This is caused by the heat capacity of the soil being much higher than that of the air. Nevertheless, Rettig (1956) described the principle that demonstrates the influence of the air temperature on the temperature via the changes in the water vapour pressure gradient in the soil. This occurs in the event of a long-lasting drop in the air temperature. The positive water vapour pressure gradient causes a large loss of latent heat by evaporation. Accordingly, Hanks et al. (1967) explain the growth of the drop of soil temperature in dependence of increasing soil moisture content and wind speed. However, Cellier et al. (1996) point out that with an increase in the wind speed even the temperature of dry soil decreases. This is caused by an increase in the coefficient of atmospheric resistance. Therefore, soil temperature is indirectly affected by such complex and remote mechanisms as the thermal stability of the atmosphere.

The atmospheric factors affecting soil temperature also include precipitation, the influence of which on soil temperature is systematically described in Table 1 in the "soil properties" category. Brooks and Rhoades (1953) described in detail the drop in soil temperature after rainfall as a result of increased evaporation, which had already been found by Ramdas and Dravid (1934). Another mechanism affecting soil temperature within the context of precipitation is the heat exchange by water convection, simultaneously with the conduction of water between rainwater and the soil environment. More recently, a significant drop in soil temperature after heavy showers was also observed by Vysoudil et al. (2012).

INTERPRETATION AND SIMULATION OF SOIL TEMPERATURE: THE PROBLEM OF SPATIAL VARIABILITY

As early as 1980 Philip (1980) mentioned that the main problem when measuring the physical properties of the soil and their spatial interpretation is the deterministic heterogeneity of the environment. More specifically, Shein et al. (2009) dealt with the problems of the variability in soil temperature at the level of polypedons and concluded that under high requirements for precision, the measurement is representative only for a particular pedon. If the measurement results are to be interpreted in a horizontal context (spatial gradients, interpolation), at least two simultaneous measurements must be carried out at a particular tessera (Buchan 2001). According to Oke (2006), in environments with strong anthropogenic impacts, such as a city, it is necessary to place even more sensors in one location to obtain representative information on soil temperature. The spatial heterogeneity of the factors affecting soil temperature and the complexity of their relationships indicate the complexity of the interpretation and simulation of the temperature characteristics of the soil. Despite the development of computing capabilities and the significant advances made in the creation of dynamic 3D models in meteorology and hydrology, the spatial interpretation and the modelling of the soil temperature regime is still a very problematic task. Muerth (2008) divides the wide range of models simulating soil temperatures into physical-deterministic models and empirical-statistical models. Both approaches, however, often overlap and complement each other, which gives rise to "hybrid" models. In Table 2, the approaches to the simulation of soil temperature are divided in more detail into physical-deterministic models based on the heat equation, statistical-empirical models, hybrid models, multicomponent physically deterministic models and comprehensive physical-deterministic system models, which are a matter for the future.

Physical-deterministic approaches

The physical-deterministic models are based on the laws of heat conduction in the soil published by Fourier (1822). Van Wijk and de Vries (1963) presented an analytical solution of the heat conduction equation for periodic changes in the soil surface temperatures in homogeneous isotropic conditions. Knowing the two boundary conditions (the daily amplitude at the soil surface and the average temperature of the soil at the depth of onstant annual temperature), we can define the soil temperature regime at any depth. After a surface temperature and soil thermal diffusivity have been supplied, the analytical solution of the heat equation gives rise to a sinusoidal model of the soil temperature regime (e.g. Elias et al. 2004). Lei et al. (2011) tried to overcome the robustness and rigidity of the models based on the analytical solution of the heat equation. Using the asymmetric function (a negative energy balance has a shorter duration in the daily regime of soil temperature), the flexible relationship of the ground air temperature and a variable taking into account the soil moisture in the calculation of the amplitude damping coefficient, they presented a powerful model with an acceptable amount of input data. The input data consists only of the ground air temperature and soil moisture (thermal diffusivity is simply included within the variable that represents the soil moisture). Therefore, this model does not consider the composition and structure of the soil.

The methods used for the numerical solution of the heat equation have also attracted the attention of researchers. It was primarily the coupled

Table 2 Schematic representation of the strengths (+) and weaknesses (−) of different approaches to soil temperature simulations.

Time axis	Soil temperature simulation method	Demand on input data	Accuracy	Validity in space	Practical usage
19c	Physical-deterministic models based on the heat equation	++	+	− −	− −
	Physical-deterministic models based on coupled heat transfer equation	− −	+++	− − −	− − −
	Statistical-empirical models	+++	− −	−	−
	Hybrid models	−	+	−	+
	Multicomponent physical and deterministic models	− −	+	+	+
21c	Comprehensive physical-deterministic system models	− − −	++	++	+++

heat transfer equation presented by Philip and de Vries (1957) which inspired many scientists to use a numerical solution of the heat conduction and convection in the soil (e.g. Warren and Price 1961; Thomas and He 1995; Li and Sun 2008).

The concept of the coupled heat transfer equation, i.e. effective thermal conductivity, is methodologically the most exact way of simulating the soil temperature regimes. In practice, however, it is hardly usable because of the high demands regarding the input data. Therefore, many other works have appeared that attempt to work out in detail the determination of the spatiotemporal distribution of the „ordinary“ thermal conductivity of soils, using submodels based on empirical relationships, which gives rise to hybrid models. For example, Tarnawski and Leong (2000) tried to implement the critical value of soil moisture (critical in the sense of a significant break of thermal conductivity) into selected models of thermal conductivity in two ways: through the dependence of the thermal conductivity on the clay content and using a fraction with a permanent wilting point.

The problem with the practical use of the majority of models based on the solution of the thermal conductivity equation is the need for detailed spatial data otherwise they are valid only in situ. The other drawback of these models is a lack of sensitivity to the effects of meteorological conditions on soil temperature. These limits were appreciated by Johansen (1977), who came to the conclusion that it is beneficial also to create empirical models that can be compared and gradually build a parallel to purely analytical solutions. By means of this approach, he opened a space for the creation of various empirical-statistical models.

Empirical-statistical approaches

Many of the empirical-statistical models focus on the possibility of simulating soil temperature using the air temperature. Roodenburg (1985) attempted to determine soil temperature 10 cm below the surface on the basis of a function of the course of the temperature throughout the year and the minimum and maximum air temperatures of the previous day. Zheng et al. (1993) compiled a simple and relatively accurate empirical model simulating soil temperature on the basis of 11 daily moving averages (originally for a depth of 10 cm). Variants on this model simulate soil temperatures under bare soil, soil shaded by vegetation cover and soil covered with snow. The solid accuracy and simplicity of this model were confirmed by Kang et al. (2000) and Lehnert (2013).

Wu and Nofziger (1999) conclude that, when using the air temperature as the input edge condition of models, it is suitable to increase the maximum and minimum temperatures by about 2°C, while for soils shaded with vegetation this increase should be lower. On the other hand, Muerth et al. (2012) pointed out that the relationship between soil temperature and the ground air temperature is more variable over time than expected.

Radiation characteristics (duration/total of global radiation, surface net radiation) are used less frequently for the simulation of soil temperatures. Oliver et al. (1987) consider the estimate of the heat flows in the soil based on net radiation to be unreliable, pointing out the variability of active surface properties. Their statement is confirmed by the results of the correlation analysis between soil temperature and global radiation carried out by Valuš (1974). However, if the net surface radiation is accompanied by an assessment of the flow of heat into the subsoil, the influence of the active surface can be avoided using an empirical constant (Muerth 2008). Mihalakakou (2002) presented a model calculating the impacts of the weather that works on the principle of a neural network and found out that if only data on global radiation intensity was used, the correlation coefficient between the measured and estimated soil temperatures reached a higher value than that when only data on the air temperature was used. Consequently, we can assume that the potential of global radiation for the simulation of soil temperature has not yet been fully exploited. However, according to Mihalakakou (2002) soil temperature correlates better with the relative humidity than with any other meteorological characteristics.

Current multi-component high-capacity models

As a result of the low density of the research-point network, the interpretation and prediction of the soil temperature regime in a larger area has been a schematic issue until recently. With the advancement of geospatial technologies and remote sensing of the soil, the development of complex models simulating the energy flows in the landscape has recently seen significant progress. The first comprehensive models to simulate soil temperature over a wider area have also appeared. Their characteristic feature is an attempt to achieve an exact physical-deterministic approach.

The first example of these multi-component models is the TERRA_LM model of the German Meteorological Service. TERRA_LM is based on the analytical solution of the heat conduction equation using a phase (multi-layered) method and is a part of a wider project of climate modelling on a small scale. It takes into account the snow cover, transpiration, soil moisture and other climatic elements in the spatial resolution of a 7×7 km grid (Schrodin and Heise 2001). The German Meteorological Service has so far published only a simplified version of the model, which expects sandy loam soil for the whole of Germany. However, the TERRA_LM model shows very good results and is continually being expanded and refined (Schrodin, Ritter and Heise 2006).

Another example of a multi-component large-volume model is SHTM (Muerth 2008). The SHTM model created for the upper Danube basin effectively combines physical and geographical approaches and takes advantage of the fact that there is a relatively dense network of well-equipped meteorological/climatological stations. Like the TERRA_LM model, the SHTM model is methodologically based on the phase method of thermal conductivity. The limiting factor of this model is the application of less detailed soil maps for the supply of the physical properties of the soil (1 : 1 000 000 or 1 : 200 000). The Canadian NEST model (Zhang et al. 2003, 2005) is even more robust than the TERRA_LM and SHTM models. The NEST model was applied at a resolution of 0.5° north latitude/west longitude throughout the whole territory of Canada. Like the TERRA_LM and SHTM models, the NEST model uses a one-dimensional heat equation and the so-called phase (multi-layer) method for the heat transport in the soil. The edge conditions of the model equation are set to a depth of 35 cm below the earth's surface and on the upper level of vegetation. Therefore, this model is theoretically coherent, but it imposes high demands on the amount of input data and depends on partial submodels. The bottom layer comprising the geological subsoil requires data on geothermal energy (this is resolved by a submodel). The pedosphere is divided into three horizontal layers with a specific thermal conductivity on the basis of the interpolation of data from a soil survey. In the upper layer of the pedosphere, the organic content is also taken into account (again by using the submodel). The snow dynamics also requires a submodel. In addition, the radiation and energy balance of the surface (the interaction between the layers) is calculated. Another layer, the ground layer of the atmosphere, requires knowledge of the meteorological conditions (with the application of submodels of air temperature, pressure, water vapour, precipitation and global radiation). The influence of vegetation in the highest layer of the model is resolved using the modified land cover category, along with the calculation of the leaf area index. Although the authors acknowledge that soil temperature can vary significantly within the grid, in the places where soil temperature measurements took place, a tight correlation was achieved between the measured and simulated soil temperatures at a depth of 20 cm.

DIRECTIONS OF FURTHER RESEARCH

There is a wide range of works describing the effects of many factors on soil temperature. The findings gained, however, were mostly approached without a comprehensive view of the environment in which they were obtained. Considering the high level of spatial heterogeneity and complexity in the relationships of factors affecting soil temperature, in most cases, the conclusions gained are valid only for the area of interest and thus do not reach the desired universal validity. Therefore, the limiting factors for the further development of the study of soil temperature include: 1) methodological inconsistencies in the study of soil temperature and factors affecting soil temperature and 2) the low density of points with a representative measurement of soil temperature. With the research in this condition, despite the considerable progress of computing capabilities, the potential for the application and validation of current multi-component large-volume models is significantly reduced.

In more recent works, the pursuit of the unification and extension of the documentation of the factors affecting soil temperature in the neighbourhood of research areas is obvious. The WMO (2008) suggests that, for the purposes of the monitoring of soil temperature, the description of the research area should include at least the soil type, surface character, gradient and slope orientation. Shein et al. (2009) state that research leading to a definition of dynamic fields of soil temperature should be preceded by a spatial exploration of the soil surface, analysis of the basic physical properties of the soil and the creation of schematic soil maps, a map of the isopleths of the physical properties of the soil at a given depth, the collection of samples and laboratory determination of the major hydro-physical and thermo-physical properties of the soil. Such a procedure is, however, financially and technically demanding and time-consuming, and therefore hardly workable over a larger area.

Soil temperature can be seen as a state function of an open subsystem of the geosphere, or as a dynamic property of the pedosphere, whose features, properties and processes are mutually conditioned and repeat in time and space with some regularity and similarity. This is confirmed by the findings of Pierson and Wight (1991), who analysed many factors and concluded that if there is a time cyclicality in the soil temperature regime, it should also be possible to find a spatial cyclicality. Therefore, understanding soil temperature in the context of geotopes on the basis of a process-oriented approach in the sense of Mosimann (1990) can be a good starting point for further studies.

CONCLUSIONS

Despite the fact that there are a large number of works dealing with the effects of various factors on soil temperature, it is not possible to use this knowledge universally to refine the spatial simulation and prediction of soil temperature. This is due to the high spatial-temporal heterogeneity and complexity of the interactions of various factors. Therefore, the studies of the impacts of these factors on soil temperature should continue with a primary focus on a systematic approach – for example, within the context of an elemental geosystem. As a result of the development of computer capabilities and geostatistic devices, the main limiting factor in the development of the study of soil temperatures is the unsystematic approach to the study and the lack of relevant spatial data for the application and validation of complex multi-component models, rather than the complexity and interdependence of the processes determining soil temperature.

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Résumé

Faktory ovlivňující teplotu půdy jako limity prostorové interpretace a simulace teploty půdy

Teplota půdy ovlivňuje průběh řady významných procesů v celé krajinné sféře. Její studium přesto patří k poměrně málo frekventovaným výzkumným tématům. Navzdory výraznému technickému a metodickému pokroku ve studiu procesů probíhajících krajinné sféře zůstává při interpretaci a predikci prostorové variability teploty půdy velká míra vědecké nejistoty.

Doposavad jen málo uspořádané poznatky o faktorech působících na teplotu půdy byly konceptuálně rozděleny na faktory vlastností půdního prostředí, faktory charakteru georeliéfu a aktivního povrchu a faktory stavu atmosféry. Napříč všemi zmíněnými kategoriemi se přitom prolíná vliv lidské činnosti. Poznatky o vlivech vlastností půdního prostředí na teplotu půdy, mezi které patří především chemické složení, zrnitostní složení, pórovitost a půdní vlhkost, vychází z četných prací zabývajících se spíše termickými vlastnostmi půdy. Vlivy charakteru georeliéfu a aktivního povrchu na teplotu půdy je naopak nezbytné odvozovat především z modifikací energetické a inherentní radiační bilance. Hlavními faktory ovlivňující teplotu půdy skrze radiační bilanci povrchu půdy jsou vedle intenzity radiace zejména orientace povrchu, charakter povrchu, albedo a zastínění povrchu okolním reliéfem. V rámci energetické bilance je nezbytné uvažovat také latentní tok tepla. Protože v reálné krajině je velká část povrchu půdy pokryta vegetací, ukazuje se, že nelze opomíjet také problematiku transmitance, absorpce a transformace záření na povrchu rostlin, fyziologické procesy rostlin a následný vznik specifických mikroklimat porostů. Vlivy charakteru georeliéfu a aktivního povrchu na teplotu půdy jsou proto vždy provázány s možnými projevy místní modifikace klimatu (stavu atmosféry) na teplotu půdy. Za nejdůležitější atmosférické faktory vzhledem k teplotě půdy jsou považovány procesy transformace, disperze a reflexe záření v atmosféře. Přímá závislost mezi teplotou půdy a intenzitou záření je však jen malá. Další vlivy meteorologických (klimatických) faktorů na teplotu půdy zůstávají skryty

v mechanismech přeměny energie na povrchu půdy a šíření tepla v půdě. Opomíjet nelze především působení srážkové vody, rychlosti větru a parciálního napětí vodních par v atmosféře.

Vzhledem k vysoké prostorové heterogenitě a složitosti vzájemných vztahů faktorů působících na teplotu půdy je modelování a simulace teploty půdy velmi náročným výzkumným tématem. Současné přístupy k modelování teploty půdy byly rozděleny do pěti kategorií: fyzikálně-deterministické modely založené na rovnici vedení tepla, fyzikálně-deterministické modely založené na složených rovnicích vedení tepla, statisticko-empirické modely, hybridní modely (kombinace předchozích) a vícesložkové fyzikálně deterministické modely. Problémem praktického využití většiny modelů založených na řešení rovnice vedení je potřeba zajištění podrobných prostorových dat. Jejich dalším nedostatkem je necitlivost k vlivům atmosférických faktorů na teplotu půdy. Jako varianta k tomuto přístupu proto vznikají empiricko-statistické modely, které však nedosahují požadované prostorové přesnosti a univerzálnosti. Snaha minimalizovat nevýhody fyzikálně-deterministických a statisticko-empirických modelů vyústila v tvorbu hybridních modelů. Toto řešení však není metodicky ucelené. V poslední době se tak pozornost výzkumu obrací k vícesložkovým velkoprostorovým modelům, které díky modernímu geostatistickému aparátu kalkulují se širokou škálou faktorů. Lze proto předpokládat, že v budoucnosti budou vznikat první komplexní systémové fyzikálně-deterministické modely vyžadující rozsáhlou datovou základnu. Takové modely bude možné považovat za zcela novou kategorii přístupů k modelování teploty půdy.

Přes velké množství prací, které se vlivy jednotlivých faktorů na teplotu půdy zabývají, se doposavad nedaří získané poznatky efektivně využívat k zpřesnění prostorové simulace a predikce teploty půdy. Příčinou je vysoká časoprostorová heterogenita a složitost vzájemných interakcí jednotlivých faktorů. Působení faktorů na teplotu půdy by proto mělo být nadále studováno především systematicky – např. v kontextu procesně orientovaného elementárního geosystému. Pro rozvoj studia teploty půdy není vzhledem k rozvoji výpočetní techniky a geostatistického aparátu nadále hlavním limitujícím faktorem samotná složitost a komplexnost procesů determinujících půdní teplotu, ale nesystematičnost přístupu ke studiu teploty půdy a s ní související nedostatek relevantních prostorových dat.

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