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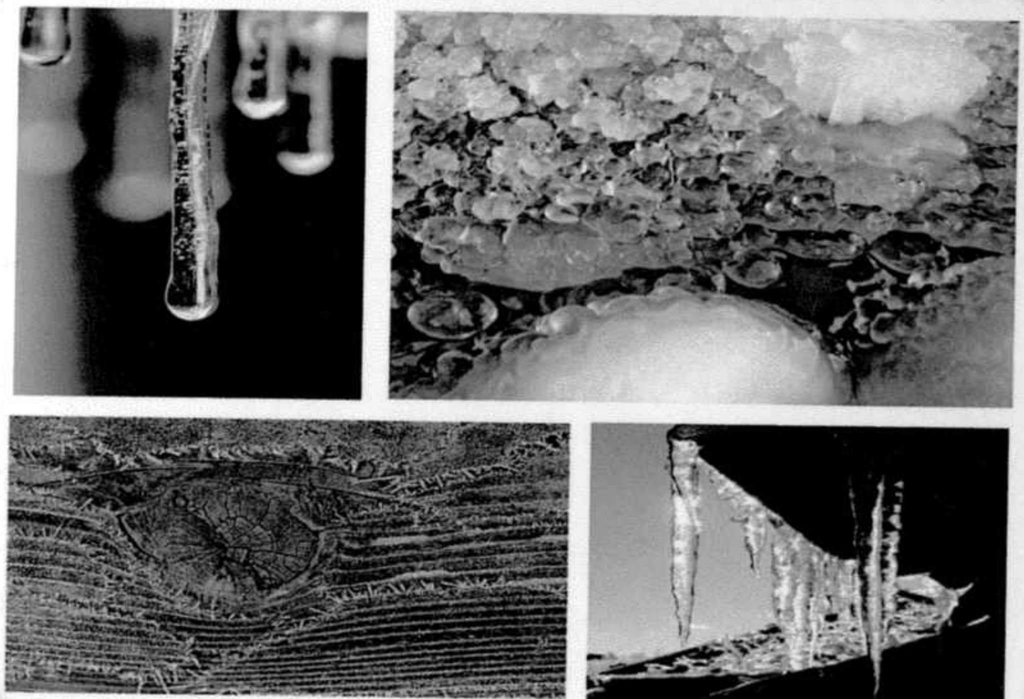

Max I. Krugger & Harry P. Stern
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Chapter 2

PERMAFROST MODELING IN WEATHER FORECASTS AND CLIMATE PROJECTIONS

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ABSTRACT

This chapter briefly reviews current state-of-the-art in modeling permafrost in numerical weather prediction models (NWPMs), chemistry transport models (CTMs) and in general circulation models (GCMs) and earth system models (ESMs) for projecting the global climate. Pros and cons of various methods are assessed. Deficits of GCM/ESMs permafrost modeling practice are discussed based on gridded observed soil-temperature data; deficits of the treatment of permafrost in NWPMs and CTMs are elucidated by examples of site-by-site evaluations. In addition, the uncertainty in simulated soil moisture and heat fluxes due to uncertainty in soil physical and plant-physiological parameters is illustrated. The consequences of incorrect simulation of or even neglecting of permafrost processes for simulated weather and climate are discussed. Extreme changes in permafrost distribution and active layer depths, as they are associated with wildfires/fires and their impact on the simulated atmospheric conditions, are addressed as well. Finally the great challenges for improving permafrost simulations (grid resolution, lack of horizontally and vertically high resolved soil data, uncertainty in soil parameters, organic soils) in GCMs, ESMs, CTMs and NWPMs and how to address these challenges is outlined.

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INTRODUCTION

At high latitudes and at high altitude of mountainous terrain, permafrost (defined as soil in which temperatures remains continuously at or below freezing point for, at least, two consecutive years) and the active layer (which thaws seasonally) are the primary subsurface components of the land-atmosphere system. Permafrost restricts the mobility of soil-water, and infiltration. Thus, capillary action, infiltration, and percolation are rather inefficient in permafrost. An important aspect of permafrost is the local temporal equilibrium between the ice, gaseous, and liquid phase of water within the soil. Any changes in heat diffusion and conduction caused by a change in snow thickness, insolation at the soil surface or infiltration affect all three water phases in the soil and soil temperature simultaneously. Any change in soil temperature results in freezing, thawing, sublimation or water vapor deposition and a release of latent heat or consumption of energy, again altering soil state variables (soil temperature, soil volumetric liquid water, ice and water vapor content) and fluxes (e.g., soil heat flux, soil water flux, soil water vapor flux). Thus, freeze-thaw cycles affect the thermal and hydrological properties of soil because of the release of latent heat and consumption of energy accompanied with phase transition processes.

Ice changes the dynamics of soil thermal fluxes through the dependence of soil volumetric heat capacity and thermal conductivity on soil volumetric water and ice content. The specific heat capacity of water is twice that of ice and the thermal conductivity of ice exceeds that of water about four times.

Regions with permafrost are characterized by low winter air-temperatures, low saturation pressure of water vapor and frequently stable stratification of the atmospheric surface layer. All these conditions lead to low evaporation. Consequently, soil moisture remains stored in the frozen active layer. Thus, in spring the capacity of the soil to take up additional water will be limited even if some of the soil already thawed. In consequence of all these, this means that permafrost may enhance spring peak flood events (Cherkauer and Lettenmaier 1999). In summer, transpiration and evaporation depend on the active layer depth, soil and vegetation type as well as on meteorological conditions.

The hydrological and thermal surface conditions associated with permafrost and the active layer affect the near-surface atmosphere, and hence weather and climate, by the exchange of heat, moisture, and matter at the soil-atmosphere interface (e.g., Stendel and Christensen 2002, Mölders and Walsh 2004). At the same time, the active layer depth is sensitive to weather; on the long-term, permafrost temperature and stability and active layer depth are sensitive to climatic change (e.g., Kane et al., 1991, Lawrence et al., 2006, Mölders and Romanovsky 2006). Permafrost thawing not only can cause huge economic and infrastructure damages and ecosystem changes, it also releases water and trace gases; the related changes in trace gas and water cycle and ecosystems again can feedback to climate. (e.g., Esch and Osterkamp 1990, Cherkauer and Lettenmaier 1999, Oechel et al., 2000, Serreze et al. 2000, Romanovsky and Osterkamp 2001, Zhuang et al., 2001). The coupling between soil moisture and thermal processes is fundamental to high-latitude soil irritations. Therefore, this coupling has to be considered appropriately in numerical weather prediction models (NWPMs) to capture the annual soil-temperature cycle and 2m-temperatures in winter (e.g., Viterbo et al., 1999). For all these reasons permafrost, permafrost dynamics, and soil-water freezing and soil-ice melting have also to be considered in climate and earth system modeling and for climate impact assessment.

Apparently permafrost variations have yet to receive a concerted effort within the context of global climate and earth system modeling. Recently, Luo et al. (2003) examined the performance of 21 modern land-surface models (LSMs) using their standalone versions and soil temperature observations along with fluxes and snow data from the 18-year Valdai dataset, a site without permafrost, but with regularly frozen ground in winter. Their study revealed that explicit inclusion of soil-water freezing and soil-ice melting improves the prediction of soil temperature and its seasonal and inter-annual variability. To appropriately represent the heat, moisture, and matter exchange at the soil-atmosphere interface, modern NWPMs, General Circulation Models (GCMs) and Earth System Models (ESMs) require suitable LSMs to simulate frozen ground and permafrost dynamics. For GCMs and ESMs such LSMs are indispensable for investigations of permafrost-climate feedbacks. ESMs also need to consider permafrost dynamics for examination on climate-permafrost-ecosystem change feedbacks.

Over the last decades geologists and geophysicists have developed site-specific permafrost models for investigation of permafrost dynamics (e.g., Goodrich 1982, Nelson and Outcalt 1987, Kane et al. 1991, Romanovsky and Osterkamp 1997, Smith and Riseborough 2001, Zhuang et al. 2001, Ling and Zhang 2003). Due to their fine vertical grid increments ($\leq$0.05 m) these kinds of permafrost models usually consume huge amounts of computational time. Since investigations of permafrost dynamics are typically oriented at decades or even centuries and the geological processes associated with changes in permafrost distribution are relatively slow, these kind of models typically run at large time steps (Mölders and Romanovsky 2006). Furthermore, most permafrost models are site-specific and calibrated (e.g., Romanovsky et al. 1997, Osterkamp and Romanovsky 1999, Romanovsky and Osterkamp 2001), i.e. new data have to be collected for calibration when they are supposed to be applied elsewhere (Mölders and Romanovsky 2006). Such calibration involves that the majority of available data at a site serves to determine optimal soil-transfer parameters, leaving the rest of data for model evaluation.

Applying a typical calibration technique certainly would lead to better predictions than those that are typically obtained with soil models designed for use in NWPMs, GCMs or ESMs. However, performing such a calibration technique for these models would require consistent soil temperature data for calibration for the typical domains of NWPMs and worldwide for GCMs or ESMs. As of today no such dataset exists making usage of calibrated permafrost models in NWPMs, GCMs or ESMs technically impossible. Furthermore, it has yet to be determined whether calibration coefficients may be climate sensitive. Coupling a permafrost model with a NWPM, GCM or ESM would also be a challenge because NWPM, GCM or ESM simulations require input of water and energy fluxes to the atmosphere at time steps of less than a minute to several minutes; consequently any vertically highly-resolved permafrost model would have to be run with this time-step making such coupled simulations computationally unattractive and in the case of weather forecasting even prohibitive. For all these reasons, the weather forecast, climate and earth system modeling communities do without calibration. They instead have developed various physically based concepts for predicting permafrost, active layer and soil frost processes. In doing so, knowledge and well-accepted concepts from permafrost and atmospheric sciences have been combined to build suitable soil models considering frozen soil physics for use in NWPMs, GCMs and ESMs.

All modern numerical NWPMs, GCMs and ESMs apply soil models embedded in their LSMs to simulate the thermodynamic and hydrological surface forcing (e.g., temperature, specific humidity, fluxes of water vapor and sensible heat) at the soil-atmosphere interface. The atmospheric scientific community has developed these soil models based on the best knowledge, and spent great efforts to evaluate and improve them (e.g., Yang et al. 1995, Shao and Henderson-Sellers 1996, Lohmann et al. 1998, Dai et al. 2003, Mölders et al. 2003a). Incomplete knowledge of soil type and heterogeneity as well as soil initial conditions generally limits the predictability of the soil state, fluxes of heat, trace gases, water vapor, and water, and phase-transition processes within the soil and at the atmosphere-soil interface. Further errors in simulating soil conditions and fluxes result from the necessity to parameterize sub-grid scale processes, prescribe soil physical parameters, discretized partial differential equations and incorrectly simulated forcing (e.g., precipitation rate and amount, insolation).

In the following, the theory of soil physics, history and current state of modeling soil physics in atmospheric sciences is reviewed and evaluated; error sources for simulated permafrost quantities are discussed.

THEORY OF SOIL PHYSICS

The earth system has various interactions between its various spheres (lithosphere, cryosphere, atmosphere, ocean). These interactions occur at various temporal and spatial scales. Depending on the time scale one is interested in certain processes are so slow that they seem not even to exist and hence are negligible at this scale (Fig. 1). For instance, at the typical forecast range of NWPMs (up to 5 or 10 days) the spatial distribution of permafrost does not change, but the active layer depth may change notably; over a typical climate period of 30 years that is considered by ESMs or GCMs, however, the spatial distribution of permafrost may significantly (even in a statistical sense) change in response to atmospheric warming or cooling over this climate period.

In the absence of impermeable layers soil-water motion in the vertical is more distinct than lateral soil-water movements due to gravity forces. Typically lateral soil-water movement, V, is of the order of up to several centimeters per day. In any atmospheric model, the horizontal extension of model grid cells, L, is several hundred meters to several 100 kilometers. Scale analysis shows that on the typical time scale, T, of atmospheric models the lateral soil-water movement is several orders of magnitude smaller than vertical soil-water transport (see Fig. 1).

The vertical heat- and water-transfer processes and soil-water/soil-ice freezing/thawing can be expressed based on the principles of the linear thermodynamics of irreversible processes (e.g., de Groot 1951, Prigogine 1961) including the Richards-equation (e.g., Philip and de Vries 1957, Philip 1957, de Vries 1958, Kramm 1995, Kramm et al. 1996, Mölders 1999). The governing balance equations for heat and moisture including phase transition processes and water extraction by roots χ read (e.g., Philip and de Vries 1957, de Vries 1958, Sasamori 1970, Flerchinger and Saxton 1989, Kramm et al. 1994, 1996, Mölders et al. 2003a)

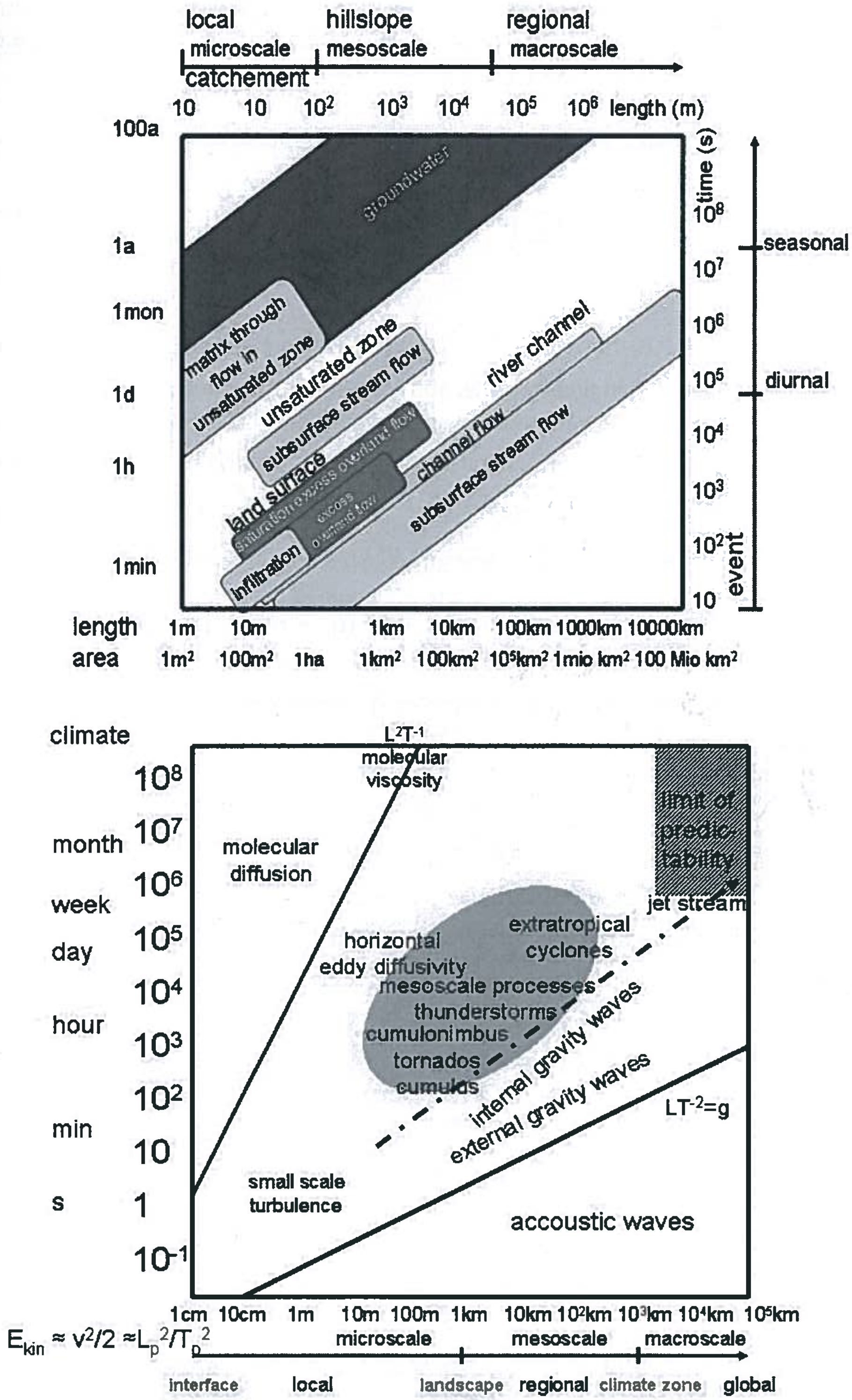


Figure. 1. Schematic views of hydrological (left) and atmospheric (right) scales modified after Bronstert et al. (2005) and Hantel (1997), respectively.

$$C\frac{\partial T_S}{\partial t} = \frac{\partial}{\partial z_S}\left(\lambda\frac{\partial T_S}{\partial z_S}\right) + \frac{\partial}{\partial z_S}\left(L_v\rho_w D_{T,v}\frac{\partial T_S}{\partial z_S}\right) + \frac{\partial}{\partial z_S}\left(L_v\rho_w D_{\eta,v}\frac{\partial \eta}{\partial z_S}\right) + L_f\rho_i\frac{\partial \eta_i}{\partial t} \tag{1}$$

$$\frac{\partial \eta}{\partial t} = \frac{\partial}{\partial z_S}\left(D_{\eta,v}\frac{\partial \eta}{\partial z_S}\right) + \frac{\partial}{\partial z_S}\left(D_{\eta,w}\frac{\partial \eta}{\partial z_S}\right) + \frac{\partial}{\partial z_S}\left(D_{T,v}\frac{\partial T_S}{\partial z_S}\right) + \frac{\partial K_w}{\partial z_S} - \frac{\chi}{\rho_w} - \frac{\rho_i}{\rho_w}\frac{\partial \eta_i}{\partial t} \tag{2}$$

Here z_S, λ, L_v, L_f, T_S, η, η_i, $D_{\eta,v}$, $D_{\eta,w}$ and $D_{T,v}$ are soil depth, thermal conductivity, latent heat of condensation and freezing, soil temperature, volumetric water and ice content, and the transfer coefficients for water vapor, water, and heat. Soil hydraulic conductivity $K_w = k_s W^{2b+3}$ depends on the saturated hydraulic conductivity k_s, pore-size distribution index b, and relative volumetric water content $W = \eta/\eta_s$ (e.g., Clapp and Hornberger 1978, Dingman 1994). The volumetric heat capacity of moist soil (Mölders et al. 2003a)

$$C = (1-\eta_s)\rho_S c_S + \eta\,\rho_w c_w + \eta_i\,\rho_i c_i + (\eta_s - \eta - \eta_i)\rho_a c_p \tag{3}$$

depends on the porosity of the non-frozen soil, η_s, the densities of dry soil, ρ_S, water, ρ_w, ice, ρ_i, and air, the specific heat of dry soil material, c_S, water, c_w, ice, c_i, and air at constant pressure. Soil volumetric heat capacity increases with increasing soil moisture for most of soils (e.g., Oke 1978). The thermal conductivity λ of unfrozen ground is a function of the soil-water potential $\psi = \psi_s W^{-b}$ also called matric potential, suction and tension head with ψ_s being the saturated water potential. Figure 2 exemplarily shows for various soil-types the dependence of thermal diffusivity on relative volumetric water content. At soil temperatures below $0°C$, thermal conductivity depends on volumetric ice and water content. Thermal diffusivity more than doubles twice when relative volumetric water content increases from 0.5 to saturation (W=1; Fig. 2).

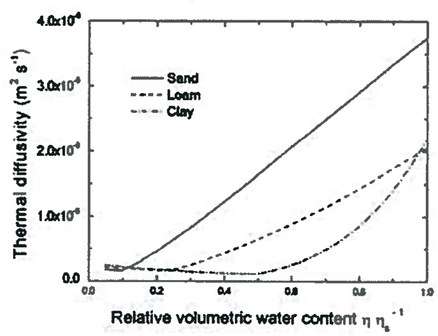


Figure. 2. Dependence of thermal diffusivity on relative volumetric water content for various soil-types. Modified from Mölders (2001).

The transfer coefficients for water vapor, water and heat are given by (Philip and de Vries 1957, Kramm 1995, Kramm et al., 1996)

$$D_{\eta,v} = -\alpha \nu D_w b \frac{\eta_s - \eta}{\eta} \frac{\rho_d}{\rho_w} \frac{g\psi}{R_d T_S} \tag{4}$$

$$D_{\eta,w} = -\frac{b k_s \psi_s}{\eta} \left(\frac{\eta}{\eta_s} \right)^{b+3} \tag{5}$$

$$D_{T,v} = \alpha \nu D_w (\eta_s - \eta) \frac{\rho_d}{\rho_w} \frac{L_v - g\psi}{R_d T_S^2} \tag{6}$$

Here g, α, ν, D_w, ρ_d, and R_d are gravity acceleration, a torsion factor that considers curvatures in the soil material due to roots (Sasamori 1970, Zdunkowski et al. 1975, Sievers et al. 1983, Kramm et al. 1996), a correction factor that is typically close to 1, the molecular diffusion coefficient of water vapor in moist air, and the density and gas constant for dry air, respectively.

In Eq. (1), the first term on the right side represents soil-temperature changes by divergence of soil-heat fluxes. The second term describes the divergence of soil-heat fluxes due to water-vapor transfer. The third term expresses how a soil-moisture gradient contributes to the soil-temperature change (Dufour effect), and the last term addresses soil-temperature changes due to freezing/thawing. In Eq. (2), the first two terms on the right side give the changes in volumetric water content caused by divergence of water vapor and water fluxes. The third term indicates how a temperature gradient contributes to the change in volumetric water content (Ludwig-Soret effect). The saturation vapor pressure is a function of soil-temperature. Consequently, a soil-temperature gradient leads to differences in saturation pressure and a water vapor flux that modifies soil moisture. This phenomenon is well known to exist in other porous media (e.g., snow). The fourth term gives changes due to hydraulic conductivity, the fifth considers water uptake by roots, and the last term represents changes due to freezing/thawing. The Ludwig-Soret and Dufour effects are cross-phenomena typically considered in the thermodynamics of irreversible processes.

If ice is present, soil-water potential Ψ, will remain in equilibrium with the vapor pressure over pure ice given by (Fuchs et al. 1978)

$$\Psi = \pi + \frac{L_f (T_S - 273.15)}{g\, T_S} \tag{7}$$

Here π is the osmotic potential. Osmotic effects due to solutes are typically omitted in NWPMs. However, they should be considered in chemistry transport models (CTMs), GCMs and ESMs in conjunction with solute chemistry because thawing of the active layer or permafrost releases traces gases (e.g., methane).

At any given soil temperature below 0°C all water in excess of (Flerchinger and Saxton 1989)

$$\eta_{max} = \eta_s \left\{ \frac{L_f \left(T_S - 273.15\right)}{g\,\psi_s\,T_S} \right\}^{-1/b} \quad (8)$$

freezes. Figure 3 exemplarily shows the dependence of maximum liquid water content on soil temperature for some selected soil-types. Considering the differences in volumes taken by water and ice, the volumetric ice content

$$\eta_i = \left(\eta_{total} - \eta_{max}\right) \frac{\rho_i}{\rho_w} \quad (9)$$

is proportional to the difference of the total water (liquid, solid, gaseous) within the soil layer minus the maximum liquid water content for temperatures below freezing point

Water extraction by roots and the following transpiration act as a soil-water sink. Soil-water uptake by roots, among other things, depends on vegetation-type, soil-physical and geologic characteristics, plant available soil-water, soil-temperature, aeration, competition or interaction with roots of other species, fertilizer, biologic and soil-chemical processes and transpiration. Various parameterizations have been developed with varying complexity (e.g., Gardner 1960, Cowan 1965, Federer 1979, Sellers et al. 1986, Martin 1990, Mölders et al., 2003a). The main differences between the various approaches are the assumptions on water-uptake restrictions, root-length, vertical distribution, whether or not root distribution varies with time, soil and/or vegetation type. Most recent LSMs used in atmospheric models assume equal distribution of roots in the root zone or only distinguish between the upper and lower root space (cf. Table 1). In the latter case, it is further assumed that the boundary between the two root spaces falls together with a soil-layer boundary; the same is true for maximum root length (e.g., Wilson et al. 1986, Martin 1990).

Simulating Frozen Ground

Since the cross-effects are very small under most conditions and since volumetric heat capacity and thermal conductivity of the substrate influence each other only marginally, decoupled equations to describe the energy- and water-transport within the soil are commonly used (e.g., Deardorff 1978, McCumber and Pielke 1981, Groß 1988, Dickinson et al. 1993, Schlünzen 1994, Jacobson and Heise 1982, Eppel et al. 1995, Chen and Dudhia 2002, Dai et al. 2003). However, this decoupling is realized in various ways as described in the following. Table 1 lists the various methods used by recent soil models of NWPMs, GCMs, and ESMs.

Table 1. Classification of soil-models used in NWPMs, GCMs and ESMs with respect to parameterizations used and model approaches. The symbol X indicates free choice of the number of layers, values in brackets are the typical choice in the models indicated.

Name	Reference	Purpose	Number of layers for			Treatment of			
			η, η_{ice}	T_s	root	moisture	temperature	roots	Soil frost
SECHIBA	Ducoudre et al. 1993, Polcher	GCM	2	2	2	Top-to-bottom filling			
NCEP	Chen et al. 1996	NWP	2	2	1		force-restore		
BATS	Dickinson et al. 1986, 1993, Yang et al. 1997	GCM	3	3	3	Richard's equation	heat diffusion	Federer-Cowan-model	Uniform freezing to –4C, limited thermal diffusivity
BEST	Cogley et al. 1990, Pitman		3	3	2	diffusion	heat diffusion		Explicitly, ice and water co-exist
BASE	Verseghy 1991, Desborough 1997, Pitman et al. 1991, Slater et al. 1998	GCM	3	3	3	Richard's equation	heat diffusion		Explicitly, ice and water co-exist
PROGMOD [1]	Ács et al. 2000	Substitutes surface flux data in diagnostic model	3	3	1	Richard's equation	force-restore	Hornet-approach, equal distributed	Bulk-heat capacity, freezing/ melting
SURF[2]	Claussen 1988	mesoscale modeling	2	5	0	Force-restore	heat diffusion	none	none
SURFW[3]	Claussen 1988, Mölders 1988, Fröhlich & Mölders 2002	mesoscale modeling	2	5	0	Force-restore	heat diffusion	none	none
HTSVS[4]	Kramm et al. 1994, 1996, Mölders et al. 2003, Mölders & Rühaak 2002, Mölders & Walsh 2004	Dry deposition, LSM in mesoscale modeling	X	X	X	Coupled diffusion equation including Richard's equation		Cowan-type, different in upper/lower root space	Explicitly, ice and water coexist

[1] Used by Universität Wien, Universität Budapest, Universität Bayreuth
[2] Used by GESIMA at GKSS, Universität Leipzig; similar LSMs are used by FITNAH at Universität Hannover or METRAS at Universität Hamburg, and Institut für Troposphärenforschung Leipzig
[3] Used by GESIMA at Universität Leipzig
[4] used by MM5 at EURAD Universität zu Köln, Universität Leipzig, (older Version at Universität Frankfurt)

Table 1. (Continued)

Name	Reference	Purpose	Number of layers for			Treatment of			
			η, η_{ice}	T_s	root	moisture	temperature	roots	Soil frost
OSULSM[5]	Chen & Dudhia 2001a,b	LSM in mesoscale modeling	X (4)	X (4)	3	Richard's equation	heat diffusion		none
MOSAIC	Koster & Suarez (1992)	GCM	3	2	2	Darcy's law	Force restore		
BUCKET	Manabe 1969, Robock et al. 1995, Schlosser et al. 2000	GCM	0	1	1	bucket	heat balance		
PLACE	Wetzel & Chang 1988, Wetzel & Boone 1995	Flexible	7	50	2	Richard's equation	heat diffusion		
SiB	Sellers et al. 1986, Xue et al. 1991, Sud & Moko 1999	GCM, mesoscale	3	2	2	diffusion	Force-restore	Federer-Cowan-model	
SEWAB[6]	Mengelkamp et al. 1999, 2001, Warrach 2001	Macroscale hydrological, mesoscale atmospheric models	6	6	1	Richard's equation	heat diffusion	different in upper/lower root space	
VEGMOD[7]	Schädler 1990, Grabe 2001								
ECHAM[8]	Dümenil & Todini 1992	LSM in GCM	1	5	2	bucket	heat diffusion	Equally distri-buted	
ECMWF[9]	Viterbo & Beljaars 1995	LSM in mesoscale modeling	4	4	3	Richard's equation	heat diffusion	Expon-ential decrease with depth	

[5] used by MM5 at EURAD Köln, LMU München, IFU Garmisch-Partenkirchen
[6] used by GKSS
[7] used by Universität Karlsruhe, Forschungszentrum Karlsruhe, IFU Garmisch-Partenkirchen
[8] used by MPI Hamburg
[9] used by ECMWF

[5] used by MM5 at EURAD Köln, LMU München, IFU Garmisch-Partenkirchen
[6] used by GKSS
[7] used by Universität Karlsruhe, Forschungszentrum Karlsruhe, IFU Garmisch-Partenkirchen
[8] used by MPI Hamburg
[9] used by ECMWF

Name	Reference	Purpose	Number of layers for			Treatment of			
			η, η_{ice}	T_s	root	moisture	temperature	roots	Soil frost
TERRA[10]	Jacobson & Heise 1982	LSM in mesoscale modeling	3	3	1	Force-restore	Force-restore	Resis-tance	no
WASIM[11]		Hydrology					none		
BIOME_BCG[12]	Running & Hunt 1993.	Ecology	1	1	1	Top-to-bottom flow (no upwards flow)	Force-restore		
LSM_FU-Berlin	Blümel 2001	Determine H from satellite data	0	0	0	-.-	-.-	-.-	
ISBA[13]	Noilhan & Planton 1989, Mahouf & Noilhan 1991, Douville et al. 1995, Boone et al. 2000	LSM in GCM, mesoscale modeling, NWPM	2	2-3	1	Force-restore	Force-restore		Explicit, ice and water coexist
SWB	Chen et al. 1996, Schaake et al. 1995	hydrology	2	-.-	1	Richard's equation	-.-		
CLASS	Verseghy et al. 1991, 1993, Chatta & le Treut 1994, de Ronag & Polcher 1998	GCM	3	3	3	Darcy's law	Heat diffusion		Linear, temperature dependent freezing/melting
AMBETI[14]	Braden 1995	Agricultural consulting	3	3	3				
GISS	Abramopoulos et al. 1988, Lynch-Stieglitz 1994	GCM	6	6			Heat diffusion		

[10] Used by LM, DM at DWD, IfT, Universität Bonn, by REMO at MPI Hamburg, TU Cottbus
[11] used by ETH Zürich
[12] Used by Universität Bayreuth, BITÖK, PIK
[13] Used by FOOT3D, Universität zu Köln, Institut für Geophysik und Meteorologie
[14] used by DWD, Universität Bayreuth

Force-Restore Method

The force-restore method had been introduced by Deardorff (1978) and became standard for NWPMs in the mid-Eighties. A force-restore model (Fig. 4) considers at least a thin top layer of depth d_1 and a deep soil layer of depth d_2 for which the soil temperature and moisture states are calculated. The force-restore model considers two distinctly different time scales in soil. The conditions in the uppermost layer are governed by the rapid responses to atmospheric forcing (e.g., precipitation, evaporation, diurnal course of atmospheric heating). These changes are represented by the so-called force term. The deeper soil layer only responses slowly to the atmospheric forcing. It typically represents annual changes. The interaction between the upper and deeper soil is considered by the restore term that describes the supply of heat and soil moisture from the deep soil layer. Some versions of the force-restore model consider a third layer that considers decadal variation. In all layers, prognostic equations are solved to determine soil temperature or moisture conditions. In doing so, soil temperature and moisture conditions are assumed to be independent from each other except if freezing/thawing is considered. Then these phase transitions lead to a change is soil temperature.

NWPMs that use the force-restore method are limited in resolving the various soil horizons (Montaldo and Albertson, 2001). High latitude soils, however, frequently show a very heterogeneous vertical stratification because they were formed by during the ice age. Moreover, the force-restore methods does not permit for simulating the vertical distributions of soil processes like the diurnal variation of the boundary between an unfrozen upper and a frozen deeper soil layer because it works with only two or three reservoirs. However, surface-water and energy fluxes are extremely difficult to predict without knowing the exact depth of the freezing line.

NWPMs that use the force-restore method are, for instance, the Deutschland Model (DM) of the German Weather Service (e.g., Jacobson and Heise, 1982), the APREGE of Météo-France and the Spanish Weather Service that both use ISBA (Noilhan and Planton 1989, Mahfouf et al., 1995); GCMs using a force-restore method are CSIRO9, and the ARPEGE climate model (DéQué et al., 1994, Mahfouf et al., 1995). See also Table 1.

Multi-Layer Models

Multi-layer soil models (Fig. 4) are most suitable for permafrost simulation in NWPMs, CTMs, GCMs and ESMs because they permit for simulating the vertical distributions of soil processes like the diurnal variation of the boundary between an unfrozen upper and a frozen deeper soil layer. Consequently, huge efforts have been spent to enlarge multi-layer soil models by soil-frost processes (e.g., Koren et al., 1999, Boone et al., 2000, Warrach et al. 2001, Mölders et al. 2003a, Narapusetty and Mölders 2005, 2006). Koren et al. (1999), for instance, tested and evaluated a soil-frost model offline that now is included with modifications in the NCEP (National Center for Environmental Prediction) Eta model. Mölders et al. (2003a) included the physics of soil-water freezing and thawing of soil-ice into the soil-model of the Hydro-Thermodynamic Soil Vegetation Scheme (HTSVS; Kramm et al.

1994, 1996) that is used in several mesoscale meteorological models (e.g., GESIMA Mölders and Rühaak 2002; MM5 Mölders and Walsh 2004).

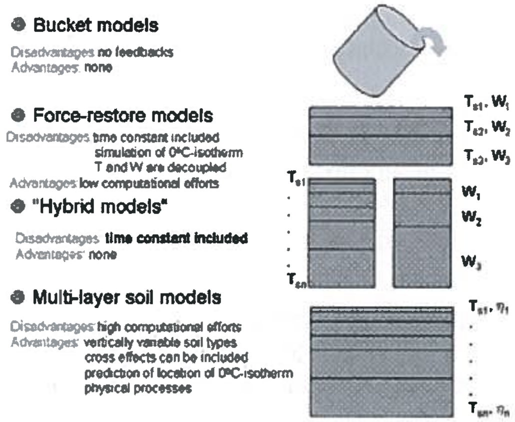


Figure 4. Schematic comparison of different concepts used for soil modeling in atmospheric models.

The impacts of soil-water freezing and soil-ice thawing in the active layer and the related processes have received little systematic study in the context of their influence on short-term weather.

Obviously, the coupled equation set (1) and (2) includes cross-effects like the Dufour effect (i.e., a moisture gradient contributes to the heat flux and alters soil temperature) and Ludwig-Soret effect (i.e., a temperature gradient contributes to the water flux and changes soil volumetric water content). Such a set of equation has either to be solved simultaneously by an iteration technique or must be simplified to avoid the iteration required by the coupling due to the cross-effects. Typically the interactions between the soil thermal and moisture regimes by the Ludwig-Soret and Dufour effect are neglected because they are negligible small under many circumstances. These interactions become noteworthy when chemicals are considered, for which they should be considered in CTMs and ESMs, when soil conditions suddenly switch from the dry to the wet mode, when soil temperatures vary around the freezing point, during snow-melt, and over the long-term these processes may gain influence on other processes or variables (Mölders and Walsh 2004). The Dufor-effect, for instance, was found to affect soil temperature up to 2 K, the Ludwig-Soret effect affects water recharge by 5 % of the total recharge over the long-term (Mölders et al., 2003b). Changes in soil temperatures and moisture caused by these cross-effects may alter the exchange of heat and moisture at the atmosphere-soil interface under these conditions.

The partial differential equations have to be discretized by a numerical scheme. Typically in LSMs of CTMs, NWPMs, GCMs and ESMs the Crank-Nicholson-scheme sometimes in conjunction with Gauß-Seidel-techniques are used (e.g., Kramm 1995). When using a Crank-Nicholson-scheme it is advantageous to introduce a logarithmic coordinate transformation into Eqs. (1) and (2) by $\xi = \beta \ln(z/z_D)$ before integrating to apply equal spacing and central differences for well appropriate finite difference solutions. Here, z_D is the lower boundary, and β is a constant which is to be chosen for convenience. Sensitivity studies showed that discretizing the partial differential equations by a type of Galerkin finite element scheme is advantageous for simulation of frozen soil physics (Narapusetty and Mölders 2006).

In most LSMs of NWPMS and CTMs thermal conductivity is assumed to be either constant or parameterized by using McCumber and Pielke's (1981) empirical formula (see also Kramm 1995, Kramm et al., 1996)

$$\lambda = \begin{cases} 419\exp(-(P_f + 2.7)) & P_f < 5.1 \\ 0.172 & P_f \geq 5.1 \end{cases} \tag{10}$$

With and $P_f = 2 + {}^{10}\log|\psi|$. Many state-of-the-art LSMs of NWPMS, CTMs, GCMs and ECMs use this parameterizations or variations thereof. For soil-temperatures below the freezing point the LSMs of many NWPMs, CTMs, GCMs and ESMs assume a mass-weighted thermal conductivity depending on the amounts of liquid and solid volumetric water content present

$$\lambda = \frac{\lambda_w \eta + \lambda_i \eta_i}{\eta + \eta_i} \tag{11}$$

Here the indices w and i stand for the liquid and solid phase of soil-water. In doing so, either a fixed or calculated value of thermal conductivity for the liquid and a value of 2.31 J/(msK) or so is used for the solid phase.

Predicted soil-temperature is highly sensitive to the thermal conductivity of the soil. Mölders and Romanovsky (2006) showed that the parameterization of thermal conductivity according to Eq. (4) provides much higher thermal conductivity values than typically found for permafrost soils; Eq. (4) also provides a decrease of thermal conductivity as the ground freezes, while observations typically indicate the opposite effect. In permafrost, thermal conductivity can be determined as Farouki (1981)

$$\lambda = \lambda_s^{(1-\eta_s)} \lambda_w^{(\eta_s - \eta_i)} \lambda_i^{\eta_i} \tag{12}$$

This formula is often applied in permafrost modeling (e.g., Lachenbruch et al., 1982, Riseborough 2002). Here, λ_s, λ_w(=0.57 W/(mK)), and λ_i(=2.31 W/(mK)) are the thermal conductivity of dry soil, water, and ice, respectively. Typical values for λ_s range between 0.06 and 0.25 W/(mK) (e.g., Pielke, 1984). In permafrost soils, typical values for λ range between 0.7 and 2.4 W/(mK) (e.g., Romanovsky and Osterkamp 2000).

Since permafrost soil pores are typically totally ice-filled, Farouki's formulation does not consider the possibility of partially air-filled pores because permafrost soils are usually saturated. Freezing of soil-water, however, also frequently occurs in mid-latitude winter or deserts where soil-pores are often partially filled with air. Since in NWPMs, GCMs and ESMs have also to be able to predict soil-temperatures accurately under these conditions, Mölders and Romanovsky (2006) enlarged the parameterization to include the impact of air

$$\lambda = \lambda_s^{(1-\eta_s)} \lambda_w^{\eta} \lambda_i^{\eta_i} \lambda_a^{(\eta_s - \eta - \eta_i)} \tag{13}$$

Here λ_a(=0.025 W/(mK)) is the thermal conductivity of air. This formulation is consistent with Eqs. (1) to (3) that explicitly consider water vapor fluxes (third and first on the right side of Eqs. (1) and (2), respectively) and air (last term of Eq. (3)). It leads to Farouki's formulation in the case of permafrost soils that are usually saturated meaning (Hinkel et al., 2001) $\eta_{air} = 0$, $\eta_s - \eta_i = \eta$, and $\lambda_a^{(\eta_s-\eta-\eta_i)} = \lambda_a^0 = 1$.

The empirical formulation with mass-weighted thermal conductivity values generally provides greater thermal conductivity values than Mölders and Romanovsky's (2006) parameterizations (e.g., Fig. 5, their Fig. 3). These authors report that thermal conductivity calculated with Eq. (10) ranges between 0.292 and 5.745 W/(mK), while values of about 2.2 W/(mK) and 1.5 W/(mK) were observed in the deeper and upper soil. Using the modified version of Farouki's formula yields thermal conductivity values between 0.149 (uppermost layer after dry episode) and 1.52 W/(mK) with about 1.1 W/(mK) on average. Note that an uncertainty analysis using Gaussian error propagating techniques identified Eq. (10) as a critical source of errors in predicted soil temperature because the natural variance in empirical parameters (pore-size distribution index, saturated water potential, porosity) propagates to great uncertainty in calculated thermal conductivity (Mölders et al., 2005). Uncertainty in parameters propagates less strongly when using the modified Farouki formula, for which parameter-caused statistical uncertainty in calculated thermal conductivity, soil temperatures, and soil-heat fluxes is lower than when using the mass-weighted formulation.

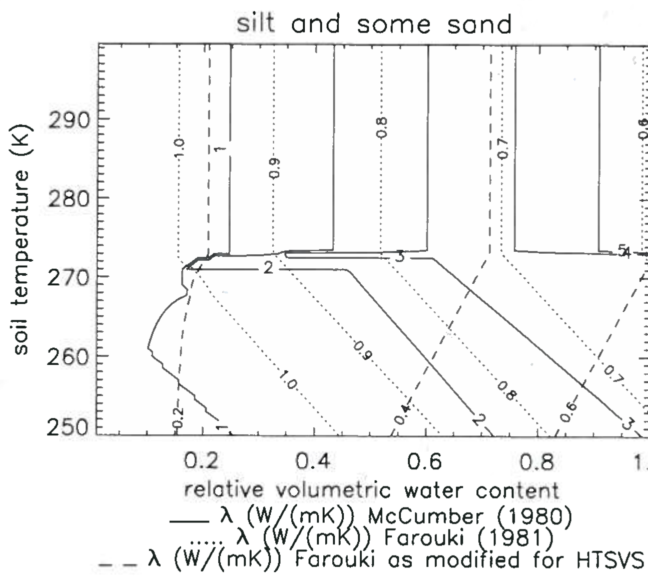


Figure. 5. Thermal conductivity as obtained by Eqs. (10) and (12). Figures for other soil types show similar basic pattern

Since the phase transitions alter soil temperature by release or consumption of heat diagnosis of soil ice has to be solved iteratively. Typically a first-order Newton-Ralphson-technique is applied (e.g., Mölders et al., 2003a).

NWPMs using multiple-layer soil models are, for instance, MM5 (Grell et al., 1994), WRF (Skamarock et al., 2005); GCMs and ESMs with multiple soil-layers are, for instance, the Community Climate System Model (CCSM) family, the ECMWF GCM and the Canadian GCM using CLASS (see also Table 1).

Hybrid Models

In hybrid soil models (Fig. 3), soil-wetness is determined by a force-restore-method (e.g., Deardorff 1978, Groß 1988, Schlünzen 1994, Jacobson and Heise 1982), while soil heat-fluxes and soil temperatures are calculated from a one-dimensional heat-diffusion equation (e.g., Claussen 1988, Groß 1988, Schlünzen 1994, Eppel et al., 1995). In these models, soil temperature layers typically differ from the two or three reservoirs used for soil moisture determination because the heat-diffusion equation is often solved for more than two or three layers to better capture the diurnal variation of soil temperature (e.g., Fig. 3). It is obvious that when soil temperature and soil moisture are calculated at different depths, permafrost hardly can be dealt with in this kind of soil model, for which they are not further discussed.

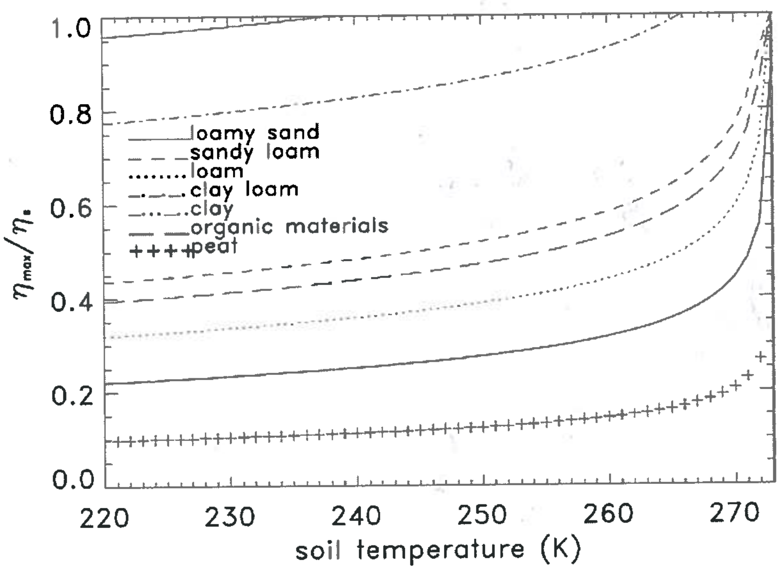


Figure. 3. Dependence of maximum liquid water content on soil temperature for some selected soil-types. From Mölders and Walsh (2004).

Vertical Resolution

In theory, fine soil-grid increments ensure accurate simulation of soil heat and moisture fluxes, temperature and moisture profiles. Unfortunately, global datasets of vertical distributions of soil type are not available. The of soil type and soil initial state data, and huge computational burden associated with a fine grid dictate the vertical grid resolution of soil models of CTMs, NWPMs, GCMs or ESMs. For reasonable turn-around times a compromise between efficiency and practical accuracy of soil-temperature is made. Modern CTMs and NWPMs typically use four to six (e.g., Smirnova et al., 1997, 2000, Chen and Dudhia 2001, Skamarock et al., 2005, Grell et al., 2005, Mölders and Kramm 2007), GCMs and ESMs about ten logarithmically spaced soil layers that cover a depth down to 2 to 3m (e.g., Bonan et al., 2002, Stendel and Christensen 2002, Dai et al., 2003). Obviously, the number and position of grid nodes plays a role in how accurately the active layer depth can be captures (Fig. 6).

Boundary Conditions

The Earth's surface is the only physical boundary condition in atmospheric models. While the soil surface is part of the lower boundary with respect to the atmosphere, it is the upper boundary with respect to the soil. The lower boundary of any soil, i.e. the bottom of a soil model, is an artificial one. Ideally, it is put at a level of nearly constant soil temperature and moisture in 20 or 30 m depths or so. Doing so is especially important in permafrost soils, where decadal soil temperature variations exist even below 15 m depth (e.g., Romanovsky et al., 1997, Mölders and Romanovsky 2006). Most modern soil models used in atmospheric models have the lower boundary around 2 or 3 m depth (e.g., Kramm et al., 1995, Smirnova et al., 1997, 2000, Chen and Dudhia 2001, Dai et al., 2003, Mölders and Walsh 2004). NWPMs and CTMs typically assume climatologic soil temperatures that vary monthly and spatially at the bottom of the soil model for at least a month. Doing so introduces artificial sources and sinks for heat and moisture (e.g., Stendel and Christensen 2002). A constant soil temperature, for instance, will act as a heat source (sink) if the actual temperature is lower (higher) at that depth. While this shortcoming may be of minor impact when regarded over the short integration times of NWPMs and CTMs and if the soil temperature is appropriately set (e.g., Narapusetty and Mölders 2005), errors may accumulate over the long integration times (of at least 30 years=climate period) of GCMs and ESMs (Mölders and Romanovsky 2006). Therefore, soil models of GCMs and ESMs usually assume constant soil moisture and heat fluxes at their lower boundary (e.g., Dai et al., 2003, Oleson et al., 2004). Most soil models of GCMs or ESMs assume zero-flux conditions at the lower boundary (e.g., Oleson et al., 2004, Nicolsky et al., 2007). However, various observations (e.g., Zhang et al., 1996, Romanovsky et al., 1997, Mölders et al., 2003a, b) show non-zero heat and moisture fluxes at 2 or 3 m, the depth typically used as the lower boundary in soil models of GCMs or ESMs. Therefore, Mölders and Romanovsky (2006) performed simulations assuming zero-flux at 30 m depth where this assumption is generally fulfilled (see also Nicolsky et al., 2007). They found that a logarithmic grid-spacing with at least 20 layers is required to appropriately capture the diurnal cycle in the active layer (see also Fig. 6), the depth of the active layer, the annual soil temperature cycle, and the timing of thawing and freeze-up.

Heterogeneity Of Soil

Data from lysimeters filled with natural soil cores taken at the same site show evidence that the natural heterogeneity of soils may lead to notable differences in ground water recharge even on relatively short-term (Mölders et al., 2003a, b). It is obvious that such differences also impact soil-moisture and temperature condition. Such heterogeneity, however, is of subgrid-scale with respect to any soil model, and hence, not considered.

Other heterogeneity stems from the spatial variability of soils. Typically grid-cells of NWPMs and CTMs cover areas of several square-kilometers, while those of GCMs or ESMs encompass several 100 square-kilometers. Obviously soil type may vary or be even different over these areas. In most NWPMs and CTMs, the soil-type dominating within a grid-cell is assumed to be the representative one for the soil conditions within that grid-cell. This means soil temperature and moisture as well as heat and moisture fluxes are calculated using the soil parameters of the dominating soil. It also means that in areas of discontinuous permafrost either permafrost soil or no permafrost soil is assumed in a grid-cell. In both cases the neglecting of heterogeneity may lead to great errors in predicted soil temperature and hence active layer depth (see Fig. 7).

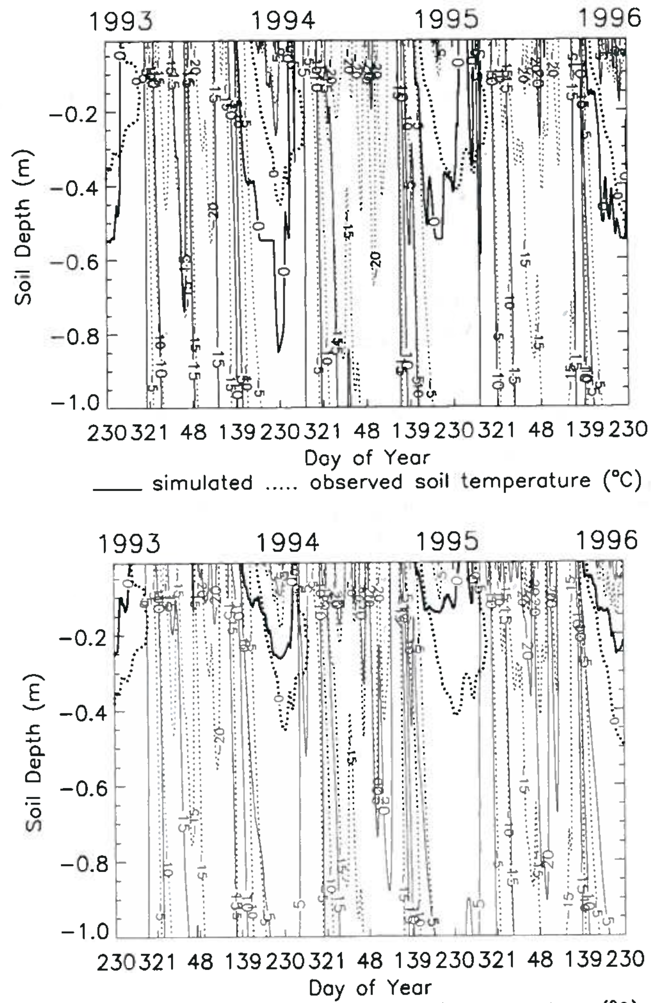


Figure. 6. Comparison of soil temperature as simulated at Barrow, Alaska with HTSVS and observed.In (a) 20 layers and in (b) 13 layers are logarithmically spaced to a depth of 30 m. Modified from Mölders and Romanovsky (2006).

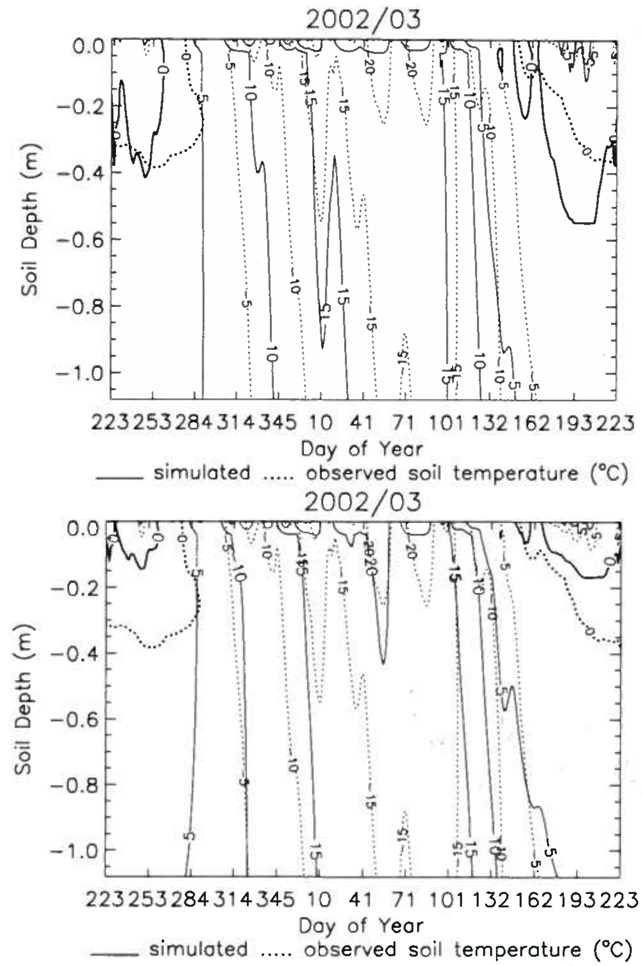


Figure. 7. Comparison of soil temperature as simulated with HTSVS and observed at Yakutsk, Siberia. In (a) soil type varies with depth, while in (b) a constant soil type, namely that of the uppermost layer is assumed. Observation data from Levine (2007; pers. communication

Some research meteorological models consider subgrid-scale spatial heterogeneity of soils by some kind of mosaic approach or subgrid-scheme (e.g., Mölders and Raabe 1996, Mölders et. al 1996). Considering subgrid-scale heterogeneity of soils can lead several Kelvin differences in soil temperature as compared to the strategy of dominant soil-type.

In modern GCMs and ESMs, the fact that soil type may vary horizontally in space is typically considered by some kind of mosaic or TOPMODEL approach (e.g., Dai et al., 2003, Essery et al., 2003, Oleson et al., 2005, Nui et al., 2005). Herein soil temperature and moisture conditions are determined for the various horizontal patches of different soil-type. The grid-cell soil temperature and moisture are then derived as an area-weighted average of the soil-temperatures of the various patches within the grid-cell.

Most modern soil models of NWPMs, GCMs and ESMs assume one soil-type for the entire soil column (e.g., Slater et al., 1998, Schlosser et al., 2000). Typically the uppermost soil-type is chosen to be representative for the entire soil column. The main reason is the lack of 3D-soil characteristic data. Nevertheless, some soil models of NWPMs (e.g., HTSVS in MM5) permit the user to consider vertical heterogeneity of soil for process research studies, rather than for general use in forecasts. Many soil models of modern GCMs or ESMs also are designed for consideration of vertically differing soil types, but basically make no use of the possibility due to the lack of 3D global distributions of soil-data. Examinations show that simulations without consideration of vertically varying soil characteristics miss many details in soil temperature and moisture patterns that result from the vertical profile of soil parameters (see Fig. 7). Mölders and Romanovsky (2006) found that even in the uppermost layer where the soil-type is the same, RMSEs between simulated and observed soil temperatures increased on average up to 0.3 K as compared to simulations with consideration of a vertically varying soil characteristic profile; moreover, simulations ignoring vertical soil characteristic profiles yield significantly different soil temperature variance than those considering it. Neglecting vertical soil characteristic profiles may yield to errors in predicting active layer depth and the timing of thawing and freeze-up of the active layer (e.g., Fig. 7).

Initialization Problem

One major problem is the initialization of soil moisture and temperature in NWPMs and CTMs. Unfortunately, global datasets of vertical distributions of soil temperature and moisture conditions do not exist.

In NWPMs and CTMs, usually the soil moisture and temperatures states obtained from the previous forecast are used as initial values for the following forecast. This procedure violates the assumption used to simplify the equations, namely that horizontal heat and moisture fluxes within the soil are negligibly small. In nature as in the model, mountains usually receive more precipitation than valleys (e.g., Müller et al., 1995). In nature, runoff on the short-term and lateral soil water fluxes on the long-term lead to moister valleys than mountains (except for glaciers where water is stored in the solid phase). Consequently, when initializing NWPMs and CTMs as described before the neglecting of lateral soil water fluxes leads to too high soil moisture in mountainous regions and too low moisture in the lower elevated terrain. In weather forecasts, these errors yield to incorrect prediction of local recycling of previous precipitation and hence wrong forecasts of convection, showers, and thunderstorms (e.g., Mölders and Rühaak 2002). In permafrost regions, some of the permafrost exists in the valleys and is fed by runoff from the mountains, i.e. in such cases cannot be appropriately captures due to the initialization method. These errors can be avoided by either inclusion of horizontal moisture transport (3D soil model), or coupling/using the soil model with a hydrological model (e.g., Mölders and Raabe 1997, Mölders et al., 1999, Mölders 2001, Walko et al., 2000, Mölders and Rühaak 2002).

Being aware that lateral soil-water movements may be important on longer time scales, in the nineties several authors (e.g., Kuhl and Miller 1992, Marengo et al., 1994, Miller et al., 1994, Sausen et al., 1994, Hagemann and Dümenil 1998) introduced parameterizations of different complexity to consider runoff in GCMs. Some kind of TOPMODEL-approach (e.g., Beven and Kirkby 1979) considers soil moisture heterogeneity (e.g., Dai et al., 2003, Essery et al., 2003, Nui et al., 2005) and permits also for consideration of discontinuous permafrost. GCMs and ESMs typically initialize soil temperature and moisture states homogenously worldwide and run the soil model with the forcing of one year for several centuries until an equilibrium soil state distribution is established.

UNCERTAINTY IN PERMAFROST MODELING

The atmospheric science community spend huge efforts on investigating uncertainty in modeling the soil conditions in atmospheric models because errors in simulated soil states and fluxes may propagate into errors in atmospheric state variables and fluxes. The following sources of error have been identified:

- Discretization, vertical and temporal resolution
- Initial and boundary condition
- Subgrid-scale heteorgeneity
- Forcing data
- Assumptions and/or parameterization concepts
- Uncertainty in soil physical parameters
- Data on soil type distribution

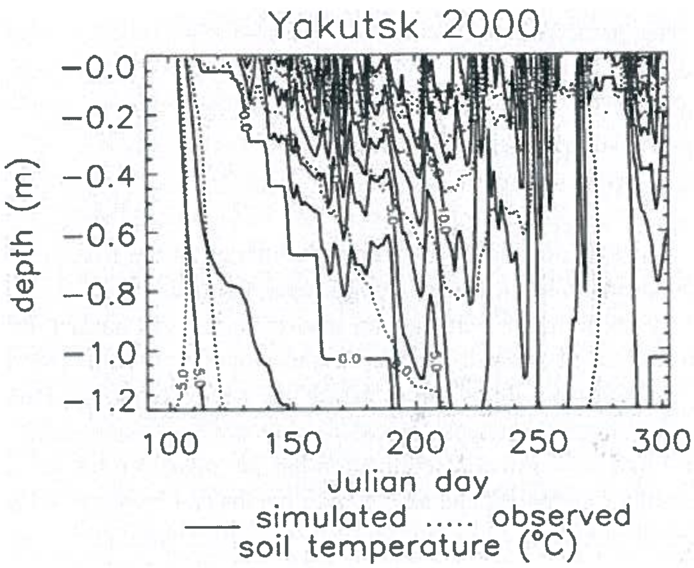
For a further discussion of the error sources mentioned in the first three bullets see also the respective subsections of section Simulating Frozen Ground.

Input of heat by precipitation, changes in insolation the soil surface due to cloudiness, changes in soil heat flux at the soil surface due to changes in wind speed can affect soil temperature, soil moisture, as well as soil moisture and heat fluxes (e.g., PaiMazumder et al., 2008). Since these changes in meteorological forcing occur on very short time scales, the temporal resolution like the vertical discretization has an impact on the accuracy with which diurnal change of soil temperatures and active layer depths can be predicted by soil models of atmospheric models. Figure 8 exemplarily shows results from simulations with different time steps and illustrates how temporal resolution can affect simulated soil temperature profiles on the long-term.

First of all, uncertainties in simulating soil temperature regimes may results from incorrectly simulated processes in the NWPM, CTM, GCM or ESM itself (e.g., Avissar and Pielke 1989, Calder et al., 1995, Mölders et al., 1996, 1997, Niu and Yang 2004).

Various sensitivity studies aimed at detecting error sources related to assumptions and/or parameterization concepts (e.g., Robock et al., 1995, Cuenca et al., 1996, Shao and Irannejad 1999). As aforementioned, the force-restore method, for instance, has only limited ability to resolve soil horizons (see Fig. 4) and to simulate the vertical distributions of soil conditions and processes (e.g., diurnal variation of the freezing line).

In any soil model in NWPMs, CTMs, GCMs and ESMs prescribed soil parameters (e.g., Table 2) represent different soil types. Ideally, the soil characteristics should be mapped as vertical and horizontal three-dimensional continuous distribution to capture the gradients and mixtures in soil type within a grid-cell or a patch of same soil type within a grid-cell. However, soils are spatially heterogeneous for which attributing a single soil type to an area or patch of several square-kilometers as it is required in atmospheric models can be ambiguous, and is a potential error source. Using a wrong soil type, for instance, can cause errors in predicted near-surface air temperatures and humidity of more than 0.5K and 0.5g/kg even in a 24-hour simulation (Mölders 2001). Assigning soil physical parameters to an area or patch is also ambiguous in GCMs or ESMs because soil surface properties can vary in time due to various events (e.g., burning of organic soils during wildfires, land avalanches, flooding, volcanic eruptions) or may be influenced by previous weather conditions (weathering) over centuries. Thus, prescribed fixed values of soil parameters for in nature time-dependent quantities may introduce uncertainty in climate and earth system modeling. Furthermore, the variability in some soil parameters is sometimes greater within the same soil type than across soil types (cf. Table 2). There is observational evidence from lysimeter studies that the heterogeneity within the same soil may cause differences in evapotranspiration and recharge of 112mm (14%) and 137mm (4%) in 5.6 years (Mölders et al.,, 2003b).



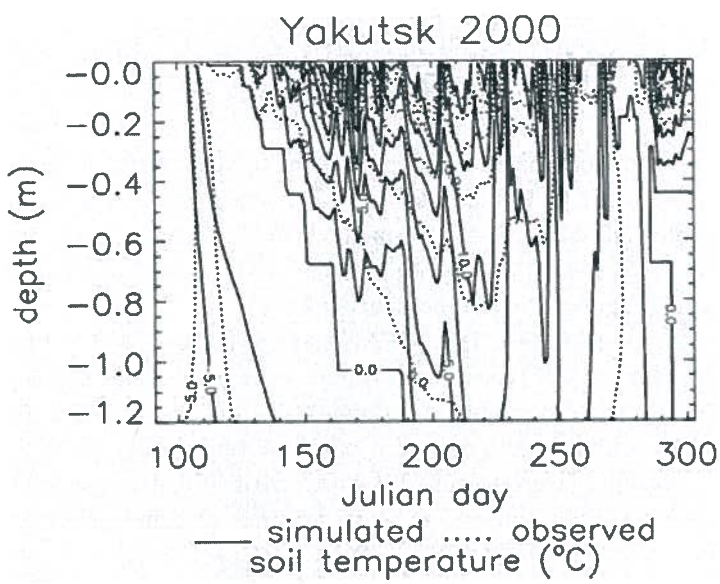


Figure. 8. Comparison of soil temperature as simulated with HTSVS and observed at Yakutsk, Siberia. The time step used in (a) is three times smaller than that used in (b). Observation data from Levine (2007; pers. communication).

Table 2. Typical mean values and standard deviations (in brackets) of soil characteristics. The symbols $k_s, \eta_s, b, \psi_s, \rho_s$ stand for the hydraulic conductivity at saturation, porosity, soil-pore distribution index, and density of the dry soil material. References are (a) Meyer et al. (1997), (b) Mohanty and Mousli (2000), (c) Schwartz et al. (2000), (e) Mendoza and Steenhuis (2003), (f) Kvaerno and Deelstra (2002), (g) Smith et al. (2003), (h) Parson (2001), (i) Wallace laboratories (2003), (j) Perfect et al. (2002), (k) Carey and Woo (1999), (l) Schlotzhauer and Price (1999), (m) Pielke (2001), (n) Grunwald et al. (2001), (o) Landsberg et al. (2003), (p) Calhoun et al. (2001), (q) Laurén and Heiskannen (1997), (r) Clapp and Hornberger (1978). Note that Cosby et al. (1984) provide slightly different values than Clapp and Hornberger (1978).

Soil-type	k_s 10^{-6} m/s	η_s m^3/m^3	b -.-	ψ_s m	ρ_s
Sand	176[r](43.9)[a]	0.395(0.056)[r]	4.05(1.78)[r]	-0.121(0.143)[r]	1580(90)[p]
Loamy sand	156.3[r](31.7)[a]	0.410(0.068)[r]	4.38(1.47)[r]	-0.090(0.124)[r]	1610(100)[n]
Sandy loam	34.1[r](13.7)[a]	0.435(0.086)[r]	4.90(1.75)[r]	-0.218(0.310)[r]	1520(140)[p]
Silt loam	7.2[r](6.2)[b]	0.485(0.059)[r]	5.30(1.96)[r]	-0.786(0.512)[r]	1400(90)[n]
Silt	2.81[r](1.325)[c]	0.476	5.33	-0.759	1420(70)[k]
Loam	7.0[r](3.028)[b]	0.451(0.078)[r]	5.39(1.87)[r]	-0.478(0.512)[r]	1350(110)[n]
Sandy clay loam	6.3[r](3.056)[e]	0.420(0.059)[r]	7.12(2.43)[r]	-0.299(0.378)[r]	1520(40)[n]
Silty clay loam	1.7[r](0.806)[f]	0.477(0.057)[r]	7.75(2.77)[r]	-0.356(0.378)[r]	1410(60)[n]
Clay loam	2.5[r](0.25)[g]	0.476(0.053)[r]	8.52(3.44)[r]	-0.630(0.510)[r]	1420(80)[n]
Sandy clay	2.2[r](8.333)[h]	0.426(0.057)[r]	10.40(1.64)[r]	-0.153(0.173)[r]	1570(120)
Silty clay	1.0[r](0.4)[j]	0.492(0.064)[r]	10.40(4.45)[r]	-0.490(0.621)[r]	1480(110)
Clay	1.3[r](0.569)[i]	0.482(0.050)[r]	11.40(3.70)[r]	-0.405(0.397)[r]	1470(140)[p]
Humus					
Peat	1.736 (0.938)[l]	0.923(0.342)	4.00(1.75)	-0.165(0.31)	106(243)
Moss	150 (400)[k]	0.900 (0.040)	1.00(1.75)	-0.120(0.310)	100(100)
Lichen	3356.5 (200)[q]	0.95 (0.060)	0.50(1.75)	-0.085(0.310)	120(30)

Various investigations using stand-alone versions of LSMs (e.g., Gao et al., 1996), NWPMs (e.g., Douville and Chauvin 2000), and GCMs (e.g., Wang and Kumar 1998) showed that initializing soil-moisture and temperature distributions is a huge source for errors in predicting the soil conditions correctly. Adjoint models and data-assimilation techniques can be applied for minimizing errors in initial soil conditions (e.g., van den Hurk et al., 1997, Callies et al., 1998, Reichle et al., 2001). Using this technique, however, is not possible for NWPMs, CTMs, GCMs or ESMs initialization due to lack of spatially continuous data.

The Project for Intercomparison of Land Surface Parameterization Schemes (PILPS) showed that LSMs strongly differ in accuracy because of, among other things, the choice of empirical parameters needed in parameterizations (e.g., Shao and Henderson-Sellers 1996, Slater et al., 1998). Typically, soil properties within a grid-cell or patch are expressed by assigning a mean value derived from laboratory or/and field studies thereby ignoring any variability. Consequently, predicted soil state variables and fluxes can differ over wide ranges in dependence of the parameter choice. Various parameter variation studies to assess whether slightly different parameters result in significant perturbations of soil temperature and moisture states. Such parameter-variation studies are subject to parameter interaction meaning that the parameter choice also affects simulated quantities that do not directly depend on the parameter. This fact makes optimal parameter choice difficult. Henderson-Sellers (1993), for instance, by using factorial experiments found that porosity is one of the most ecologically important parameters. Enhancing thermal diffusivities or volumetric heat capacities, for instance, may cool the soil and atmospheric boundary layer (locally more than 5 K and 1 K, respectively); enhancing volumetric heat capacities or thermal diffusivities may also affect atmospheric variables especially specific humidity, cloud and precipitation particles and may result in decreased maximum precipitation (Mölders 2001). Errors may also stem from incorrectly assigned soil types. An about 5 % change in soil-type distribution may alter daily averages of the soil-moisture fraction by 29 % with respect to the reference case, and surface temperature by 2.3 K (Mölders et al., 1997).

Besides systematic errors due to parameter choice, initialization, discretization, assumptions and physical parameterizations stochastic error is a source of uncertainty in predicted soil state variables and fluxes. As pointed out before, for describing soil heat and moisture transfer processes parameters have to be assigned that represent the soil characteristics. Herein stochastic errors result from the fact that the mean values of empirical soil parameters are in "error" by the amount of the standard deviation related to the natural (random) variability (Mölders et al., 2005). For many soil parameters, this variability expressed, for instance, by the standard deviation is of the same order of magnitude as the parameter itself (cf. Table 2). Consequently, any soil state variable or flux predicted with these parameters is "error"-burdened too. Such uncertainty may even reduce the trust in predicting permafrost dynamics in GCMs and ESMs. For NWPMs and CTMs, it may limit the ability to simulate the evolution of active layer depth which is important information for agricultural purposes and assessment of river runoff. For GCMs and ESMs, this uncertainty may complicate climate impact assessment.

Errors in soil state variables and fluxes related to parameter uncertainty are of random kind for which they can be evaluated with statistical methods, for instance, Gaussian error-propagation (GEP) principles. This method permits researchers to investigate the relative importance of soil physical parameters (e.g., porosity) in producing prediction uncertainty at various potential conditions. Using GEP Mölders et al., (2005), for instance, found that predicted distributions of soil temperature are less sensitive to uncertainty in thermal parameters than to uncertainty in hydraulic parameters. According to GEP results uncertainty in predicted soil-heat fluxes is within of the range as the typical errors in soil-heat flux measurements. They also found that the absolute value of soil-heat flux and its relative error decreases with increasing relative volumetric water content and concluded that soil-heat fluxes can be predicted with greater certainty after rain events or in the Tropics than under dry conditions or in dry regions.

Note that GEP can also be applied to examine how terms in the soil heat and moisture equations contribute to uncertainty in predicted soil temperature and moisture states. During phase transitions, the freeze-thaw term, for instance, can cause great uncertainty in volumetric water content and soil temperature (e.g., Mölders et al., 2005). Similar was found using other methods by Mölders and Romanovsky (2006).

SIMULATING WILDFIRE IMPACTS

Wildfires are a regular thread in many regions on Earth, so also to areas underlain by permafrost. Often the uppermost layer of permafrost contains huge amounts of organic material or completely consists of organic material like peat, moss or lichen (e.g., Beringer et al., 2001). Wildfires can burn this organic material. The degree to which this material is burned depends, among other things, on fire intensity, fire duration, and total soil water content of the material. Fires heat the soil and huge amounts of soil water evaporate during the fire. In permafrost, fire-induced changes in soil temperature go along with changes in total soil water content and the partitioning of the water phases (e.g., Hinzman et al., 2003). Consequently, infiltration, soil volumetric heat capacity and hydraulic conductivity before and after a fire differ appreciably. As compared to pre-fire soil conditions, post-fire soils are warmer. Such modified hydro-thermodynamic states of soil remain detectable long after the fire events. Due to their impact for soil temperature regime, active layer depth and soil surface temperature on the short and long-term it would be important to consider the impact of wildfire on soil temperature in the soil models of atmospheric models.

Currently, the impact of wildfires on permafrost is not considered in routine weather forecasts, CTMs, GCMs or ESMs. In CTMs, wildfire impacts on permafrost are currently neglected even when the CTMs are applied for wildfire smoke forecasts in areas underlain by permafrost. The neglecting wildfire impacts on permafrost in ESMs is despite some ESMs consider random aerosol release from wildfires and wildfire related land-cover changes in the biogeochemical cycles. One application that considered the impact of wildfires by land-cover changes and soil-temperature and moisture changes was performed with an NWPM by Mölders and Kramm (2007). Their results showed that the relatively warmer burned areas may increase atmospheric buoyancy and hence locally convection.

CHALLENGES

The lack of horizontally and vertically high resolved soil data for organic and mineral soil, uncertainty in soil parameters, and organic soils are among the biggest challenges in modeling permafrost in atmospheric applications.

Due to the lack of 3D data on the soil type distribution most modern soil models used in NWPMs, CTMs, GCMs or ESMs assume one soil-type – typically that of the uppermost soil - for the entire soil column (e.g., Slater et al., 1998, Schlosser et al., 2000). Investigations show that this simplification/assumption results in missing many details in predicted soil-temperature patterns that result from the vertical soil type profile (Fig. 7). Due to neglecting vertical soil type profile characteristics the variance of simulated and observed temporal evolution of soil temperature can differ significantly with consequences for predicted active layer depth that may be dislocated about ±0.4 m or so (e.g., Mölders and Romanovsky 2006). Off-line simulations (i.e., the feedback processes between the atmosphere and surface are not considered) with different soil hydraulic models that were run with and without ensuring consistent soil hydraulic parameters, demonstrated that uncertainty in soil hydraulic parameters overwhelms that in the theory of soil hydraulic models (Shao and Irannjad, 1999).

EVALUATION

Evaluating soil temperature and moisture conditions simulated by NWPMs, ESMs, or GCMs has been a high priority of the third PILPS phase (e.g., Henderson-Sellers et al., 1995). PILPS demonstrated that results obtained from LSMs coupled to GCMs differ on the same order of magnitude as off-line PILPS experiments; differences in LSM complexity may cause statistically significant differences in temperature, pressure, and turbulent fluxes over land (e.g., Sato et al., 1989, Thompson and Pollard 1995, Yang et al., 1995, Qu and Henderson-Sellers 1998). The results of PILPS also suggested that a soil model must be able to capture soil-temperature conditions well when run offline with observed atmospheric forcing and known site-specific parameters (necessary condition), and it must be re-evaluated when being implemented in a NWPM, GCM or ESM (sufficient condition).

Soil models of NWPMs are typically evaluated by assuming that the soil temperature and moisture measurements at a site are representative for the grid-cell within which the site is located (e.g., Chen and Dudhia 2002, Narapusetty and Mölders 2005). It is well known that some discrepancies may arise due to the fact that the model grid-cell represents a volume-average condition for several square-kilometers of several centimeters thickness. For GCMs or ESMs, however, simulated soil temperature and moisture states represent even larger volumes. Due to the large area of several 100 square-kilometers covered by GCM or ESM grid-cells often several sites exist within the same grid-cell. Thus, a comparison like performed for NWPMs becomes highly ambiguous. Therefore, GCM and ESM simulations of soil regimes are typically evaluated using gridded climatologies that are derived from point observations projected on to a grid by some kind of interpolation methods (e.g., Li 2007, PaiMazumder et al., 2008).

Recently, the digital versions of the Ground Ice Conditions map and the International Permafrost Association (IPA) Circum-Arctic Map of Permafrost (known as IPA map) were combined with ancillary data sets of the Global Land One-kilometer Base Elevation data base and the global land-cover characteristics data base to provide a gridded distribution of northern hemispheric permafrost and ground ice (Zhang et al., 2000). Such gridded data can serve for evaluation of 20th century simulations of GCMs and ESMs. This dataset, however, does not contain soil temperature or moisture conditions.

Any gridded data sets bear some uncertainty from various sources. First the data stem from routine monitoring that typically has less accuracy than specialized field campaigns. Furthermore, these data have been collected for other reasons than evaluation of GCMs or ESMs. Thus, the monitoring networks may not be representative for the landscape that a GCMs or ESMs is to cover. For evaluation of CCSM3 simulated soil-temperature climatologies in Siberia PaiMazumder et al. (2008) used gridded data based on over 400 agricultural monitoring sites. They found December-biases in soil-temperature climatology for CCSM3 of up to 6 K at 0.2 m depth of which they could explain about 2.5 K by incorrect simulated atmospheric forcing. It is obvious that the soil conditions represented by the gridded data derived there from are biased with respect to the conditions in well drained, fertile soils with other density than non-plowed soils. Moreover, agriculture is typically made on soils that have a relative deep active layer depth. Thus, great care is needed in interpreting simulated soil conditions when using these kinds of gridded data. Investigations by PaiMazumder and Mölders (2008) showed that such bias in representing the soil distribution can lead to overestimation of soil-temperature amplitudes of more than 1 K and difficulties in capturing the phase. These findings also suggest that some of the discrepancies found for GCM or ESM soil temperature simulations may be explained the networks on which the gridded climatologies are based. Taking the errors resulting from incorrect forcing and the gridded data into account, only about 1.5 K of the bias found by PaiMazumder et al. (2008) for the CCSM3 soil-temperature simulations may stem from model deficits or other error sources.

FUTURE DIRECTIONS

As discussed in the previous sections, permafrost modeling in NWPMs, CTMs, GCMs and ESMs still has several short-comings. Some of them may be addressed easily as available computer resources increase with the next generations of supercomputers, while others require serious research and data collection efforts.

Increased computational power will permit us to consider more layers and locate the lower boundary of NWPM, CTM, GCM and ESM soil models at deeper levels, i.e. reduce uncertainty related to the choice of the lower boundary condition. This way a greater depth of the soil model does not compromise the required fine resolution in the upper soil that is required to capture the diurnal and seasonal cycle of active layer depth.

As has been shown by Narapusetty and Mölders (2006) finite element schemes permit us to better capture the phase and amplitude of soil temperature variations and hence the active layer depth. Currently the computational burden is too high to run GCMs and ESMs for several decades using such methods. Therefore this improvement has to be postponed until the next generations of supercomputer will become available.

The difficulties related to initialization of soil moisture and temperature in NWPMs and CTMs could be addressed by developing a kind of analysis procedure like applied to initialize the atmosphere in NWPMs. Such an analysis method would require to measure worldwide soil temperature and moisture at the same universal coordinated time (UTC) several times a day like is common practice for meteorological data. These soil data would have to be reported and collected at a central place in an agreed upon format like GRIB that is used by the World Meteorological Organization (WMO) for reporting the huge amount of meteorological data. Some kind of interpolation procedure would have to be run to produce a hydro-thermodynamically consistent gridded global soil temperature and moisture dataset based on the latest observations.

Data of soil type distribution exist in various different data sources and must be gathered in a data center to derive a quality assessed and quality assured global gridded dataset.

ESMs that consider random aerosol release from wildfires and wildfire related land-cover changes in the biogeochemical cycles and CTMs that serve for wildfire smoke forecasts in boreal regions could be enlarged to also consider the impact of wildfires on permafrost. Doing so would require assuming a wildfire-related heat source at the top of the soil where currently the wildfire related aerosols are released and later land-cover is changed.

Non-representative network design, low site density, shut-down and/or adding sites to long-term monitoring networks can introduce substantial uncertainty in gridded data (e.g. PaiMazumder and Mölders 2009). Therefore it is an urgent need (1) to assess the potential influence of networks on gridded data derived there from, (2) to develop evaluation strategies for application of gridded data from "imperfect" existing long-term networks, and (3) to develop recommendation to improve existing networks and/or design better networks in the future.

To avoid errors from non-representative networks in gridded soil-temperature data some kind of data assimilation could be used. Similar to reanalysis in atmospheric sciences all available soil data plus meteorological forcing data as upper boundary condition in conjunction with physical soil modeling could be performed to provide some kind of reanalysis (e.g., Kalany et al., 1996, Uppala et al., 2005) for soil temperature. This method could consider the various soil types within a grid and the gridded dataset would provide a weighted soil temperature. The weighting would be with respect to the fractional coverage of a given soil-type within the grid area like soil temperatures are typically simulated in GCMs or ESMs.

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