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Perspectives in modelling climate-hydrology interactions

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Abstract

Various land-atmosphere coupling mechanisms exist that may lead to large-scale impacts on climate and hydrology. Some of them are still less understood and not adequately represented in state of the art climate modelling. But as the current generation of climate models enables consideration and implementation of important coupling processes, the present study provides perspectives for the modelling of relevant climate-hydrology interactions. On a more short-term perspective, these comprise anthropogenic land use and especially irrigation, which has been shown that it may even affect remote regions. On a long-term perspective, the coupling of hydrology to carbon cycle and vegetation becomes important, specifically the dynamics of permafrost and wetlands. Here, we present a review of current knowledge combined with some exemplary studies from a large-scale point of view. Therefore, we focus on climate-hydrology interactions that are relevant on scales utilized in current or forthcoming global and regional climate modelling exercises.

Keywords: Irrigation, Land atmosphere feedbacks, Land use impacts, Modelling perspectives, Permafrost and wetland dynamics.

1 Introduction

The hydrological cycle plays a prominent role within the Earth system and is crucially important to life on Earth including the human society. Thus, the current state of the hydrological cycle and its future development are key issues in environmental research. In studies of global and regional climate change, climate models are the current operational tools. Although the ability of climate models to simulate the various characteristics of the climate or Earth system has considerably improved within the past decades, gaps or large uncertainties in the representation of some specific processes still exist. Consequently there is a lot of room for improvement. In the following, we will focus on climate-hydrology interactions and provide two major perspectives for their modeling within the framework of Earth System Models (ESMs). Here, only those interactions are considered that are relevant on scales utilized in current or forthcoming global and regional climate modelling exercises.

Harding et al. (2011) give a general overview on current knowledge of the terrestrial global water cycle. Here, they consider aspects of state of the art global hydrology modeling, past and projected future hydrological change in means and extremes, as well as uncertainties in our understanding of the current global water cycle and how it will develop in the future. But in those aspects the feedback of terrestrial hydrology to the climate is mostly not considered. In order to investigate the interactions between climate and hydrology and how they may behave under climate change conditions, a coupled framework is necessary where both components are adequately represented. Strong interactions between the climate, hydrology, and land use occur (Claussen 2004; Falloon and Betts 2010). The snow–climate feedback is well known and described (e.g., Cess et al. 1991). However, feedbacks between CO2, vegetation, soil moisture, groundwater recharge, and climate are less well understood and are not well described in most climate and hydrological models.

Soil moisture controls the partitioning of the available energy into latent and sensible heat flux and conditions the amount of surface runoff. By controlling evapotranspiration, it is linking the energy, water and carbon fluxes (Koster et al. 2004; Dirmeyer et al. 2006; Seneviratne and Stöckli 2008). Seneviratne et al. (2006) stated that a northward shift of climatic regimes in Europe due to climate change will result in a new transitional climate zone between dry and wet climates with strong land–atmosphere coupling in central and eastern Europe. They specifically highlight the importance of soil-moisture–temperature feedbacks (in addition to soil-moisture–precipitation feedbacks) for future climate changes over this region. A comprehensive review on soil moisture feedbacks is given by Seneviratne et al. (2010). Their general principles are known (e.g. Koster et al. 2004, 2006; Teuling et al. 2009), even though there is still room for model improvement.

Soil moisture shows a high variability from daily to interannual timescales. An appropriate knowledge of soil moisture conditions is important for the initialization and quality of seasonal to yearly climate predictions. Fischer et al. (2007) indicated that the record breaking European heat wave in 2003 was enhanced by the large soil moisture anomalies that were caused by a large precipitation deficit together with early vegetation green-up in the months preceding the extreme summer event. Loew et al. (2009) showed that these soil moisture anomalies were observable using remote sensing sensors. Consequently, the impact of soil moisture memory on the climate is an important scientific topic (e.g. Seneviratne et al. 2006) and is addressed specifically in the BMBF project MiKlip PastLand where its value for seasonal to decadal prediction is investigated.

From the hydrological perspective, two major challenges for modelling climate-hydrology interactions have currently arisen where we will shed some light on in the following. On a more short-term perspective, these comprise anthropogenic land use and especially irrigation.

The coupling of hydrology to carbon cycle and vegetation is important on the long-term perspective, specifically the dynamics of permafrost and wetlands.

While the process of anthropogenic emissions due to fossil fuel burning is fairly well established in state-of-the-art climate model simulations, up to now, the possible impact of land use changes on the climate is mostly neglected in long-term climate simulation. Dale (1997) reviewed the literature dealing with the relationship between land-use change and climate change and concluded that in recent centuries, land-use change has had much greater effects on ecological variables than has climate change. Pielke et al. (2002) documented that land-use change impacts regional and global climate through the surface-energy budget, as well as through the carbon cycle, whereat the surface energy budget effects may be more important than the carbon-cycle effects. While this is valid for the past climate, results of Cox et al. (2000) indicated that carbon-cycle feedbacks could significantly accelerate climate change over the twenty-first century, and pointed out the necessity to consider the potentially large direct human influences on terrestrial carbon uptake through changes in land cover and land management. Changes in the land surface (vegetation, soils, water) resulting from human activities can affect the regional climate through shifts in radiation, cloudiness and surface temperature. Changes in vegetation cover affect surface energy and water balances at the regional scale, so that the impact of land use change may be very significant for the regional climate over time periods of decades or longer (Denman et al. 2007). The effects of a specific land use change on the climate depend on the surrounding environment and climate characteristics as a regional modeling study of Gao et al. (2003) over China has shown.

An extreme anthropogenic impact on the local hydrology is the practice of irrigation. Over 18% of total cultivated land is irrigated (Fischer et al. 2007); additionally, much nonagricultural land has been substantially modified by human activities. Conversion of land to agriculture not only impacts the local evaporation and hydrological response, but may also influence the distribution of rainfall and evaporative demand in the surrounding landscape as well as have remote impacts on the large-scale circulation. The latter will be considered in more detail in Sect. 2. Agriculture and urban development have increased substantially in the past century and will continue to develop in the twenty-first century. Therefore, any assessment of the world’s water resources must take into account both the direct and indirect influences of land use changes and the exploitation of the riverine system.

Earth’s climate is determined to a large extent by Greenhouse gases (GHG) in the atmosphere, which influence the radiation budget and thus the energy balance of the planet. Thus, fluxes that may change the atmospheric GHG content are of great importance in climate change research. Apart from water vapour and anthropogenic GHG, the various components of the global carbon cycle, especially CO2 and CH4, play a significant role. In recent years, estimates for the amount of carbon stored in soils have attracted more and more attention, and here especially the consideration of the vast permafrost regions increased numbers drastically (Tarnocai et al. 2009; Zimov et al. 2006; Schuur et al. 2008; McGuire et al. 2009). Permafrost, being defined as ground that is at or below zero degrees Celsius for more than two consecutive years, affects roughly one quarter of the Northern hemisphere (Brown et al. 1997). It is believed to store between 1400 and 1800 Pg of C in the upper few meters of the soil (Schuur et al. 2008), which would be twice the amount of the atmosphere’s content. The high northern latitudes are one of the critical regions of anthropogenic climate change, where the observed warming is clearly above average due to the so-called Arctic Amplification (Solomon et al. 2007; ACIA 2005). Climate model simulations project this trend to continue (Serreze and Barry 2011). The combination of the high C stocks in sub-arctic and arctic soils with the pronounced warming in the affected regions could thus lead to a positive feedback through the release of formerly trapped, ’deep-frozen’ C into the atmosphere, when near-surface permafrost thaws. For the thawed soils and their biogeochemistry, it is decisive whether dry or wet conditions predominate: Aerobic decomposition is relatively fast and leads to the release of CO2, while anaerobic decomposition is much slower and leads to the release of CH4 as the main product of the combustion of organic soil material. Therefore, not only the soil's temperature, but also its moisture status and specifically the presence of wetlands are important for the assessment of the biogeochemical response to climatic conditions, and thus should be represented in climate or Earth System models in a realistic and process-based manner. Thus, the adequate representation of permafrost hydrology is a necessary and challenging task in climate modelling, which will be considered in more detail in Sect. 3.

2 Impact of anthropogenic land use, especially irrigation, on climate

Several studies (e.g. Gordon et al. 2005; Piao et al. 2007; Rost et al. 2008a, b) demonstrated that land cover conversions and water withdrawals have already noticeably changed the partitioning of terrestrial precipitation into evapotranspiration and runoff. Gerten (2013) estimated that these direct human impacts have increased the global river discharge by about 5%, which is caused by the associated reduction in evapotranspiration. Regionally, the implications of anthropogenic land use may be much larger. Partially, even opposite effects (increased evapotranspiration, reduced runoff) may be induced by land use change (Destouni et al. 2013) or irrigation (Gerten et al. 2008).

Observations and model studies in tropical forests have shown effects of changing surface energy and water balance on the state of the atmosphere. For example, Marengo and Nobre (2001) found that the removal of vegetation led to decreases in precipitation, evapotranspiration and moisture convergence in central and northern Amazonia. Oyama and Nobre (2004) showed that the removal of vegetation in north-east Brazil would substantially decrease precipitation. Other model studies indicated that increased boreal forest reduces the effects of snow albedo and causes regional warming (Denman et al. 2007). Related to the latter, e.g., Göttel et al. (2008) investigated the influence of changed vegetations fields on the projected regional climate over the Barents Sea region in an off-line coupling experiment with the regional climate model (RCM) REMO and the dynamic vegetation model LPJ-GUESS (Sitch et al. 2003). They projected a forest ratio increase and a shift of the tree line to higher altitudes and latitudes caused by a warmer climate with longer snow-free periods and growing season lengths. The feedback effects to the climate of these changes were one order of magnitude lower than the effects of the greenhouse gas forcing. A further warming in spring could be attributed to the snow-albedo effect, while a cooling in summer was dedicated to changes of roughness length, enhanced transpiration and changes in surface albedo. A more extreme study was conducted by Bathiany et al. (2010) who investigated the effect of large scale changes in forest cover on global climate. They completely removed tropical forest within the ESM of the Max Planck Institute for Meteorology (MPI-ESM), which resulted in a simulated 0.4 K warming due to an increase in CO2 concentrations and a decrease in tropical evapotranspiration. A similar experiment for the high northern latitudes led to a global cooling of 0.25 K in case of complete deforestation and an equally large warming in case of afforestation. In both cases, the involved albedo changes (snow masking effect) are the main drivers of the temperature change.

Land use changes such as deforestation may have a substantial climate impact in areas located close to strong climatic gradients, such as tropical regions as well as arid and semi-arid regions. In this respect, Africa is one of the ‘hot spot’ areas. Taylor et al. (2002) stated that the climatic impacts of land use change in the Sahel region are likely to increase rapidly in the coming years. So far, the effect of deforestation and reduced vegetation cover associated with land use change in Africa has mainly been studied with coarse-grid (300 km resolution or coarser) global climate models in the form of time slice experiments and idealized forcing (see e.g. Feddema et al. 2005). With these coarse resolution models, effects on the local and regional climate can usually not be resolved. For this purpose, RCMs are an adequate tool, such as done by Paeth et al. (2009) for West Africa who conducted long-term transient climate change experiments with the RCM REMO at 50 km resolution over West Africa where they forced their simulations with increasing greenhouse-gas concentrations and land use changes until 2050. Their results indicate that significant future changes in the near-surface climate may be caused by land use changes.

A specific form of land use is irrigation, which can considerably affect the regional climate (Boucher et al. 2004, Lobell et al. 2009) and whose feedbacks onto rainfall (ter Maat et al. 2006) may become especially important where irrigation coincides with areas of global hotspots for land surface – atmosphere feedbacks. Koster et al. (2004) identified the Sahel zone as one of the hot spot areas for the feedback of surface soil wetness to subsequent rainfall. In this semi-arid region, irrigation is not a major agricultural practice, but an increase in dryland agriculture is possible which is sensitive to rainfall totals. A study of Taylor et al. (2002) showed that future likely changes in land-cover could result in a reduction of nearly 10% in rainfall. Another hot spot of soil moisture – precipitation coupling is located over India (Koster et al. 2004). The Indian subcontinent is one of the most intensely irrigated regions (Fig. 1a) in the world (Sacks et al. 2009), and many studies have shown the role of irrigation in modifying the local climate through feed back mechanisms (e.g. De Rosnay et al. 2003; Douglas et al. 2006; Lee et al. 2009); Douglas et al. 2009). Effects of irrigation, embedded in South Asian Summer Monsoon (SASM), affect 22% of world's population and hence play a crucial role in modifying the water resources, agriculture, economics and human mortality of the region. Therefore, this topic is covered separately in the following Sect. 2.1.

2.1 Impact of irrigation on the South Asian summer monsoon

As mentioned above, the effects of irrigation on local climate through feed back mechanisms are well known from earlier studies. Saeed et al. (2011) found that REMO is able to reproduce the general characteristics of the SASM, however over the land areas of north-west India and Pakistan, a systematic warm temperature bias of more than 5°C can be noticed (Fig. 2a, c). The too enhanced simulation of the heat low (Fig. 2) is a common systematic error that is present in many regional climate models applied over South Asia (Lucas-Picher et al. 2011). Since more than a decade, this heat low is used as an important predictor for SASM rainfall (Singh et al. 1995). The major part of this heat low region falls inside the densely irrigated Indus basin (Fig. 1b), which is the largest contiguous irrigation network in the world and its surface water is heavily manipulated by building large dams, link canals, watercourses etc. and hence resulting in modification of the amount of water in the soil (Khan et al. 2008). It is estimated that the Indus River drains only one eighth of the ~400 km^3 water that annually falls on the basin in the form of rain and snow, with the remainder used mostly for irrigation and returned to the atmosphere by evapotranspiration (ET) (Karim and Veizer 2002).

Saeed et al. (2009) applied the regional climate model REMO (Jacob et al. 2007) over South Asia at a resolution of ½ degree (~55 km) domain, forced with lateral boundary conditions obtained from the European Centre for Medium Range Weather Forecasts reanalysis (ERA40) (Uppala et al. 2005). In order to take into account the effect of irrigation, a map of areas equipped for irrigation (Fig. 1a; Siebert et al. 2005) was used in the simulation and the results for four SASM summer months June, July, August and September (JJAS) were presented. For the potentially irrigated fraction of a grid box, the soil wetness was increased to a critical value in each time step, so that potential ET can occur. In this way, it is assumed that irrigation is conducted to fulfil optimal conditions for the vegetation/crops, allowing them to transpire at a potential rate. Note that the main results will not change if water for irrigation is limited by available water from the rivers (Saeed et al. 2012).

In the REMO simulation without irrigation, the overestimation of the heat low (too high temperature, too low pressure; Fig. 2) resulted into increased differential heating between ocean and land, and therefore the overestimation of winds entering into the plains of the Indian subcontinent from the Arabian Sea. This causes a situation unfavourable for westward propagating currents from Bay of Bengal to intrude deep into western India and Pakistan. Therefore, less moisture is advected causing an underestimation of precipitation over this area as well. When irrigation is accounted for, a more realistic behaviour of the simulated climate is yielded. Figures 3 and 4 compare the changes of the REMO simulation with irrigation (Fig. 3b, d) to the reference simulation without irrigation (Fig. 3a, c). An improvement of simulated temperature and MSLP can be seen over the whole region, but statistically significant and most pronounced changes are present over Indus, Ganges (Fig. 1b) and southern India (Fig. 4a, b). For these regions, the standard REMO version simulated the largest systematic biases (Fig. 2), hence, the irrigation led to a better representation of these variables. Figure 4c indicates a significant increase in ET over the whole subcontinent region, again with largest increase over Indus and southern India.

The reduced differential heating of the land relative to the ocean leads to a reduction of the too strong westerly winds from the Arabian Sea into the Indian plains (Fig. 4d). This creates conditions favourable for monsoon depressions originating from the Bay of Bengal to intrude deep into the land up to western India and Pakistan. Saeed et al. (2009) could illustrate this behaviour for the development and movement of several monsoon depressions that brought rainfall to the western part of the Indian peninsula and that have been discussed in earlier published literature. Figure 5 shows one of these cases that occurred from 19-21 September 1991 (Mahajan et al. 1995). All these cases have in common that in the REMO simulation without irrigation, a depression forms in the Bay of Bengal, but it stops at the east coast of India and dissolves. Only in the simulation with irrigation, the depression is able to travel deep into land towards western India and Pakistan and to transport moisture and precipitation into these regions, such as it has been observed . Together with the increased local recycling of moisture due to the increased ET, this leads to an increase of precipitation over central/western India and Pakistan that is reducing the dry precipitation bias in this area.

For the same two model setups of REMO i.e. with and without the representation of irrigation, climate change simulations have been conducted over the South Asian model domain following the A1B emission scenario (Gerten et al. 2011). Figure 6 shows that for the projected 2m temperature changes (2085-2099 minus 1985-1999), REMO without irrigation projects an increase of more than 4°C in general and more than 6°C over the central Indian region. In contrast, the REMO simulation with irrigation projects much less warming, with a temperature increase ranging from 2°C to 4°C. This highlights the role of irrigation in attenuating the climate change signal over the South Asian region. Thus, it can be concluded that the irrigation performed over the 20th century may have already masked recent climate change signals over this region.

2.2 Conclusions and perspectives for the impact of irrigation on climate

The results presented in Sect. 2.1 signify the role of irrigation in effecting the local temperature, which in turn effects large-scale circulations and precipitation of the SASM. They also show the potential of irrigation for mitigating climate change effects in the SASM region. The present neglect of irrigation was the main cause of the systematic REMO model error over the heat low region in NW India and Pakistan that led to a too enhanced formation (too warm, too deep) of this heat low. The representation of irrigation has caused the removal of this bias. Similar biases in other RCMs (Lucas-Picher et al. 2011) suggest that they are also related to the missing irrigation process. Consequently, the representation of water used for irrigation in climate models is necessary for the realistic simulation of SASM circulation and associated rainfall. This, together with taking into account land use change, has also been emphasized by Gordon et al. (2005) for the global scale.

The impact of resolution on the irrigation effects upon the SASM has not been explicitly considered up to now. In a recent study of Tuinenburg et al. (2013), consistent results to those presented in Sect. 2.1 were found across an ensemble of three RCMs and one GCM. Here, the application of irrigation on a large scale led to changes in the large scale circulation, in which moisture shifted away from the Ganges plain towards the Indus basin and Pakistan. This has confirmed results found by Puma and Cook (2010) and Asharaf et al. (2012). But generally, a higher resolution leads to an improved simulation of the SASM. Kumar et al. (2013) summarized that most of the GCM studies focusing on the Indian monsoon region concluded that GCMs have difficulties in simulating the mean monsoon climate over India. Due to their coarse horizontal resolution, GCMs have limitations in simulating the complex orographic precipitation over India. Also, several RCM studies have been carried out to simulate the summer monsoon over South Asia, whereat all have reported an improvement in the simulation of SASM spatial and temporal distribution compared to coarser global models (Kumar et al. 2013).

80% of Indus basin river flows are attributed to the melt of snow and glacier. Considering the large impact of irrigation on SASM behaviour, one can assume that under global warming the changes in the timings of water inflows would shift towards earlier months, hence causing changes in cropping patterns and subsequently irrigation. As irrigation is impacting the climate change signal over the SASM, changes in irrigation patterns over the Indus basin will also affect the SASM circulation and associated rainfall under climate change. Therefore, not only irrigation itself but also changes in irrigation patterns need to be regarded for climate change studies over the SASM region.

While irrigation seems to have a positive mitigating impact on the SASM climate, the picture looks different for areas where the human consumption of water leads to drying and shrinking of surface waters (Asokan et al. 2010). Here, the associated decrease of evaporation from these surface waters counteracts the direct irrigation effect of increasing evapotranspiration in irrigated land areas. A very prominent example is the Aral Sea, which was the fourth largest lake on the globe until 1960, with a surface area of about 68,000 km². But large irrigation activities in many parts of Middle Asia were mainly responsible for the catastrophic desiccation of the Aral Sea within the last five decades (see Breckle and Gedyewa (2012) and references therein). How irrigation has affected the current climate or may affect the future climate under global warming conditions in other regions is an important subject for future studies. In this respect, a first ESM study was provided by Guimberteau et al. (2012). Irrigation also causes groundwater depletion over many areas of the globe (Döll et al. 2012). How this may affect climate and water resources is a prospect for future studies as the current knowledge on the impacts of changing groundwater on climate is limited.

3 Permafrost

Permafrost and wetlands are two focal points in the coupling of hydrology to biogeochemical processes under climate change conditions. A large part (~24%) of the northern hemisphere terrestrial land surface is underlain by permafrost (French 1990) that is mainly situated in high latitudes (Fig. 7). Here, climate warming is more pronounced than elsewhere, and is very likely to continue to do in the future according to Solomon et al. (2007). Permafrost soils build a globally relevant carbon reservoir as they store large amounts of deep-frozen organic material with high carbon contents. If permafrost thaws under global warming conditions, the stored carbon can be decomposed and released to the atmosphere as additional greenhouse gas, which will lead to a positive feedback. Consequently, relevant scientific questions are: How fast, how deep and to what temperature are permafrost soils going to thaw in the future?

In Sect. 3.1, relevant hydrological processes are described that occur in permafrost areas and that should preferably represented in models simulating interactions of permafrost hydrology with vegetation, climate and the carbon cycle. The current state of the representation of permafrost processes in ESMs is tackled in Sect. 3.2, while Sect. 3.3 deals with the specific topic of wetlands.

3.1 Basic hydrological processes in permafrost areas

Apart from climatic cold conditions, the occurrence of permafrost is largely controlled by physiographic features such as aspect, slope, and elevation. Other factors such as soil types, soil moisture, vegetation cover, and disturbances (e.g. wildfire) can also influence the distribution of permafrost (Haugen et al. 1982; Yoshikawa et al. 2002). The most basic process in permafrost areas is the seasonal melting and freezing of soil water in the presence of continuously frozen ground below a certain depth. The depth to which the soil is thawed is called the active layer. Regions that are affected by permafrost or extensive seasonal ground freezing show a specific behaviour of important hydrological variables: 1) Soil moisture is often rather high in near-surface layers, despite low precipitation rates in many regions; 2) river discharge observations display very low wintertime values; 3) surface runoff shows a steep spring peak after snowmelt that can deliver a substantial part of the annual total runoff (Swenson et al. 2012). Several reasons are responsible for these features, which will be described in the following.

Firstly, the phase change exerts a drop in liquid water content, and the freezing front can be seen as a water sink within the soil. This leads to the development of a gradient in liquid water content, and thus induces water movement towards the freezing front. This process is called cryosuction. It leads to unique characteristics of soil moisture in regions with permafrost and extensive seasonal ground freezing, namely to the increase of total soil moisture in the upper layers as well as to the development of large ice bodies in the ground (see also description of ice wedges below).

Secondly, the permeability for liquid water flow is reduced in frozen soils. This might be the main and most obvious effect frozen soil exerts on hydrology (Niu and Yang 2006). According to Staehli et al. (1999), there are two possible pathways for the flow of liquid water when soil temperature is below 0°C. Transport channels for slow water flows are provided by thin films of adsorptive and capillary water, which are still existing in liquid phase and whose amount depend mainly on soil texture type. Alternatively, fast water movement is possible through air-filled macropores Soils contain such pores through structural variations like cracks, holes and channels, e.g. from dead roots and soil inhabitants like worms etc.

Nearly impermeable soil layers can develop due to the freezing of the soil during winter and spring seasons (Koren et al. 1999) as ice bodies in the ground impede liquid water movement through blocking of the pore space (Swenson et al. 2012). Moreover, a strongly frozen soil will contain only very limited amounts of unfrozen water so that the ability of the subsurface material to conduct liquid water, i.e. the hydraulic conductivity, is decreased, yet not totally set to zero.

Frozen ground and snow cover also influence rainfall-runoff partitioning, the timing of spring runoff, and the amount of soil moisture that subsequently is available for evapotranspiration in spring and summer (Koren et al. 1999). For the infiltration of surface water into the soil, the above explained principles lead to the same general behaviour as for the hydraulic conductivity, as the infiltration process is lastly determined by the soil's ability to conduct water away from the surface. Nevertheless, infiltration can vary even more than the conductivity. When considering two areas with the same climatic conditions, several other locally variable factors are influencing infiltration:

- Soil texture.
- Orography and slope.
- Snow: thicker snow cover leads to enhanced insulation and a thus weaker freezing of the soil, and vice versa.
- Vegetation: similar effects as snow, via weaker or stronger insulating properties of the vegetation cover, and via its roots' effects on the soil structure.

These influencing factors can lead to patterned areas with surface water impeding zones in some parts, and with permeable zones, where conditions for infiltration are much more favourable, in others. In consequence, the impact of frozen soil on infiltration and on hydraulic conductivity is strongly scale-dependent (Niu and Yang 2006; Koren et al. 1999). Thus, it is important to notice that the falling below the Zero °C threshold does not lead to a complete blocking of infiltration and percolation. These characteristics complicate the implementation of cold regions' soil processes in land surface schemes for climate and Earth system modelling.

The response of the soil to freezing leads to specific variations in the annual cycle of soil hydrology. The snow melt, which is usually constrained to a very short period of sometimes less then two weeks, delivers a large water input to the land surface, which at this time of the year is still frozen. Infiltration capacity is thus low, and much of the snow melt water is channelled into surface runoff. The thawing of the active layer begins immediately upon the completion of snowmelt (Boike et al. 1998), dependent upon a number of factors including soil material, duration of snow cover, soil moisture and ice content, and convection of heat by groundwater (Woo 1986). The beginning of the thawing coincides with high surface moisture values, and ice melting in near-surface layers occurs on top of still frozen, and thus less permeable, deeper layers. Consequently, sub-surface water flows are weak, and high soil moisture values develop within the still thin thawed upper layers. Refreezing of infiltrated snow melt water also contributes to this (Swenson et al. 2012). Over the course of the warm summer season, the thawing and deepening of the active layer increase the water holding capacity of the soil, resulting in a decreasing surface water contribution during precipitation events and a steadily increasing baseflow contribution (Hinzman and Kane 1991). The latter is a lateral slow sub-surface runoff that can develop as the permafrost table forms a barrier to the deeper soil, where again water cannot easily percolate. Due to the enlarged water storage and increased baseflow, upper soil layers can also become drier in this part of the year. The autumn precipitation often coincides with the start of the freezing season, thus again high surface runoff rates are produced, yet much lower than in spring. During winter, the decreased hydraulic conductivity in frozen soils leads to the observed very low winter baseflow. Permafrost degradation due to a warming trend will likely lead to a decreasing seasonal variability of water flows (Frampton et al. 2011), whereat results of Frampton et al. (2013) show that total runoff will first increase and then decrease as the permafrost degradation progresses further to total thaw.

Apart from the above mentioned effects of the soil processes on hydrological quantities, perennially frozen ground shows some unique features, that are examples for processes that act on both long and short time scales, and that are often highly non-linear. Massive ice wedges are one of these features, which occur in permafrost dominated landscapes (French 1990). Water enters the soils through frost cracks and, through volume expansion during freeze-up, further increases the cavities in the ground. Cryosuction leads to movement of unfrozen, supercooled water towards the freezing front, and, over time, the ice body can grow to reach several meters in height and thickness. This process happens very slow and lasts many years to decades (French 1990). Ice wedges (and other ice bodies) are the reason for the often oversaturated ice contents in permafrost soils, and can be seen as long-term storages of both energy and water in the climate system.

However, the presence of ice wedges may also lead to abrupt changes that occur on the landscape scale when large ice bodies in the soil, after a period of relatively slow and constant warming, collapse in sudden events, e.g. due to intense rain events that bring high heat inputs into the ground. This might lead to considerable change in the landscape in form of severe soil subsidence, opening of channels, and coastal erosion. Belonging to these phenomena are the so-called thermokarst lakes, that develop when formerly stable permafrost thaws at the top due to perturbation (e.g. a fire event), and soil subsidence and melting ground ice lead to the formation of a lake (see, e.g., French 2007). This describes a particular process of wetland formation (see also Sect. 3.3). On the other hand, these thermokarst lakes may also drain by catastrophic outflow following lake tapping due to the expansion of adjacent basins or truncation by coastal retreat (see, e.g., Mackay 1988; Walker 1978; Romanovskii et al. 2000). Cycles of slow build-up of ice masses in the ground and relatively short-termed collapses in conjunction with the implied morphological changes have happened ever since. Yet since the atmosphere is warming, and since the atmospheric moisture transport from mid- to high northern latitudes as well as precipitation and circulation patterns are believed to change with anthropogenic climate change, these events might become more abundant in the future. This again has implications for the carbon cycle, as erosion events always bring formerly bound carbon back into the cycle.

3.2 Representation of permafrost processes in ESMs

The climate modelling community has a long history in systematic model intercomparison through the climate model intercomparison projects (CMIPs; Meehl et al. 2000). Results from CMIPs provide a good overview on the respective state of ESM model accuracy and performance. Koven et al. (2012) analysed the performance of ESMs from the most recent CMIP5 exercise over permafrost areas. They found that the CMIP5 models have a wide range of behaviours under the current climate, with many failing to agree with fundamental aspects of the observed soil thermal regime at high latitudes. This is partially related to the fact that most of these models do not include permafrost specific processes, not even the most basic process of freezing and melting of soil water. Moreover, the land surface parameterizations used in GCMs usually do not adequately resolve the soil conditions (Walsh et al. 2005), which often rely on either point measurements or on information derived from satellite data.

Although a good understanding of many permafrost-related hydrologic processes exists at the point and hillslope scales, this knowledge had not been adequately or systematically incorporated even into process based meso-scale hydrologic models (Vörösmarty et al. 1993) for a long time. Models on point/hillslope scales were generally constrained to one-dimensional domains of vertical extent only (Riseborough et al. 2008), which usually could not be up-scaled to larger scales due the complexity of physical interactions in permafrost regions. Also Bolton (2006) has identified a lack of process-based hydrology models that adequately simulate the soil moisture dynamics at the watershed scale and also include a realistic land-atmosphere exchange in permafrost dominated regions. But such models are required to bridge the gap between the point/hillslope scale understanding and the scale of RCMs and GCMs by capturing the hydrologic behaviour and variation of individual watersheds. Recent developments started to fill this gap (see ,e.g., Schramm et al. 2007; Bense et al. 2009; Frampton et al. 2011), thereby responding to the need for an increased realism of numerical permafrost models highlighted by Riseborough et al. (2008) and Woo et al. (2008).

As the development of meso-scale permafrost-related hydrology models has only gained momentum within the last years, it is not surprising that the situation is comparable or even worse for climate models. Until recently, the representation of frozen ground physics, as well as of above mentioned characteristics (Sect. 3.1) like reduced permeability of frozen soil and impeded infiltration of spring melt water, were, if at all, represented in a rather simplified way. An intercomparison study of different land surface schemes especially with respect to cold regions' climate and hydrology revealed large differences between the models, even in case the implementation of frozen ground physics was constructed in a similar way (Luo et al. 2003). Due to missing processes and related deficiencies of their land surface schemes, climate models often show substantial biases in hydrological variables over high northern latitudes (Luo et al. 2003; Swenson et al. 2012). Therefore, large efforts are ongoing to extend ESMs in this respect, in order to improve simulated soil moisture profiles and associated ice contents, river discharge, surface and sub-surface runoff. The ESM improvement over permafrost areas is, e.g., one of the research objectives of the European Union Project PAGE21 (www.page21.org).

Given the substantial range in the level of complexity and advancement of permafrost-related processes implemented in the ESMs, the large variety of results from the CMIP5 models is not surprising (see Fig. 8; Koven et al. 2012). The most comprehensive ESM land surface schemes include freezing and melting of soil water, the dependency of soil thermal properties on water and ice content, multi-layer snow schemes with snow on top of the soil instead of blending upper soil layers and snow, and the representation of soil organic matter (e.g. CLM; Lawrence et al. 2011). In contrast, many models incorporate only few of these processes. Soil hydrology is assessed using multi-layer schemes that compute vertical flows using Richards law (Richards 1931) or some of its derivations (e.g., Oleson et al. 2004), thereby replacing more and more the formerly used bucket schemes. The reduced permeability can thus be considered via coupling of soil thermodynamics and hydrology, i.e. hydraulic properties are functions of liquid soil water content only, instead of the total water content. This is refined in some models through the implementation of a freezing point depression. Reduction of infiltration at the surface is partially assessed, ranging from very simple approaches like total blocking soils in case of freezing, to the consideration of sub-grid scale variability, based on power law relationships between infiltration and the degree of soil freezing. Examples for global ESM land surface schemes that are in various states of on-going development are CLASS (Verseghy 1991), ORCHIDEE (Gouttevin et al. 2012), JSBACH (Ekici et al. 2013) and JULES (Best et al. 2011). The latter three also participate in the Page21 model improvement activities. One of the planned Page21 improvements is the development of a global scheme for the formation and drainage of thermokarst lakes that has not been implemented in any ESM up to now.

It is important to note that also for models that represent the same processes the results may diverge markedly. This can be attributed to differences in parameterization schemes and the choice of parameters, e.g. soil column depth and thickness of its layers, as well as to the choice of input data, e.g. soil porosity or heat capacity of the soil's dry material. In addition, initial conditions play a role due to the long spin-up times of model soils. These may comprise several years for liquid and frozen soil water content, several years (de Ridder 2008) up to two decades for soil temperature and several centuries to millennia for soil carbon storages (Wutzler and Reichstein 2007; Hashimoto et al. 2011). In this respect, Christensen (1999) pointed out the importance of an adequate initialization of soil temperature and soil moisture in climate modelling experiments. An inadequate initialization of these fields may lead to transient signals that have to be suppressed as much as possible in modern numerical climate experiments as climate sensitivity experiments operate with quite small signals.

3.3 Wetlands

Thawing permafrost will also contribute to the formation (French 2007) or disappearance (Smith et al. 2005) of wetlands that currently cover about 6-8% of the land surface. Note that available global wetlands observations span a range of potential wetland coverages, partially due their different wetland definitions which they are based on (see Stacke and Hagemann 2012). Despite of this, they agree on many large scale patterns that can be seen in Fig. 9 showing the ensemble mean coverage of wetlands based on four different datasets. Due to their function as water storage, the majority of research studies found that wetlands regulate river discharge, mitigate flood events and show increased evapotranspiration compared to other land cover types (Bullock and Acreman 2005). However, some exceptions to this general behaviour have been reported (van der Velde et al. 2013) where evapotranspiration is less efficient for wetlands than for other land cover types. The extension of wetlands determines the area where anoxic decomposition instead of oxic decomposition may take place. While CO2 is released under oxic conditions, the anoxic decomposition yields methane that is a far more active greenhouse gas than CO2. Here, the water level is an important factor for the wetland's biogeochemistry which results in carbon sequestration or decomposition (e.g. O'Connor et al. 2010, and references therein). Generally, an increase in wetland area will lead to an enhanced methane production. On the other hand, a decrease will reduce moisture fluxes to the atmosphere and may lead to a reduction in precipitation. Thus, their future development is of major interest in climate change studies.

While most studies identify wetlands as net carbon sinks for today's climate conditions (Bohn et al. 2007; Gorham 1991; Friborg et al. 2003), a number of studies concluded that some wetlands might turn into carbon sources in a warmer climate (St-Hilaire et al. 2010; Gorham 1991) due to higher productivity of methane releasing microbes. Several recent studies suspected wetlands to play an important role during periods of climate change (e.g. Ringeval et al. 2011; Gedney et al. 2004; Levin et al. 2000). However, the representation of the wetland's spatial extent and its temporal variations is still a weak point in today's ESMs and needs to be improved by a better simulation of their hydrological cycle (O'Connor et al. 2010; Ringeval et al. 2010).

However, even without consideration of the carbon cycle, the wetland hydrology in itself is an important key factor in the climate system. Wetlands are often related to regions with open surface water and saturated soil. Such regions have to be considered in ESMs because of their potential feedbacks to the atmosphere (Coe and Bonan 1997). The effect of open water surfaces on the energy and water balance was investigated by several modelling studies, e.g. Bonan (1995) and Mishra et al. (2010), who reported a significant impact of wetlands on the local climate. Generally, they found a cooling of the surface in wetland dominated regions due to increased evapotranspiration, as well as an increase of the latent heat flux and a decrease of the sensible heat flux. Eventually, this could result in increased precipitation rates as shown by Coe and Bonan (1997) and Krinner et al. (2012). Furthermore, wetlands interact in several ways with the hydrological cycle of their surrounding area. Most studies report wetlands to regulate river flow, mitigate flood events and recharge groundwater (Bullock and Acreman 2003). These observations are consistent with a modelling study by Mishra et al. (2010) who found decreased surface runoff in wetland dominated regions. However, the range of possible hydrological impacts of wetlands is rather large and depends strongly on additional conditions such as topography and soil properties. This is emphasized by several studies that describe different wetland impacts such as an increased effect on flood peaks and no or a discharging impact on groundwater (Bullock and Acreman 2003). All of these processes are of great interest for impact studies that investigate how climate change might affect the water storage capacities in a region or the characteristics of river flooding.

The modelling of the hydrological cycle in wetlands and their extent dynamics has motivated a large number of modelling studies. Generally, most models follow one of two main approaches for the hydrological representation of wetlands. One approach is concerned with the redistribution of soil moisture in the model grid cell. A widely used example is TOPMODEL (Beven and Kirkby 1979). In this approach a topographical index is computed that depends on the drainage of a given area routed through a point and its slope. This index is then applied to determine the position of the local water table at that point relative to the mean water table of the whole grid cell. The grid cell fraction where the sub-grid soil moisture exceeds the soil moisture storage capacity of the grid cell is then regarded as a wetland. The TOPMODEL approach was used and improved in several studies (e.g. Barling et al. 1994; Gedney et al. 2004; Bohn et al. 2007; Kleinen et al. 2012) and is able to compute changes in wetland extent as well. While this approach is an elegant solution, it has one major problem. As the wetland fraction depends on the redistribution of the mean grid cell soil moisture, it follows that there is an upper boundary for the maximum water depth and wetland fraction. For the extreme case of a grid cell with zero slope no wetland can emerge because the mean soil moisture can obviously not exceed the maximum soil moisture capacity. However, observations indicate that flat regions appear to be more suitable for wetland formation.

The second approach is the explicit modelling of surface water. In this case depressions in the topography are identified and filled with water that results from a positive water balance. On the one hand, this can be done on a continental scale (e.g. Coe 1997, 1998, 2000) but then the quality of the wetland representation is strongly limited by resolution of the model. Alternatively, regional models allow for a higher resolution but then depend strongly on detailed soil property information (e.g. Bowling and Lettenmaier 2010; Yu et al. 2006) or are calibrated for specific catchments (e.g. Bohn et al. 2007). Decharme et al. (2008, 2011) developed a global inundation model, but its focus is concentrated on the representation of floodplains.

In contrast to these sophisticated approaches Stacke and Hagemann (2012) developed a somewhat simpler hydrological scheme that represents the global distribution and extent variability of very different types of wetlands. The scheme was designed for the application in complex ESMs on global scale with medium to coarse resolutions (50 km or coarser), as the representation of surface water dynamics is – albeit important – not strongly developed in such models. The global-scale hydrological scheme of Stacke and Hagemann (2012) has been implemented into the Max Planck Institute for Meteorology Hydrology Model (MPI-HM). It solves the water balance of wetlands and estimates their extent dynamically. The extent depends on the balance of water flows in the wetlands and the slope distribution within the grid cells. In contrast to most models, this scheme is not directly calibrated against wetland extent observations. Using MPI-HM, the spatial distribution of simulated wetlands agreed well with different global observations for present climate (Fig. 10). The best results were achieved for the Northern Hemisphere where not only the wetland distribution pattern but also their extent was simulated reasonably well. However, the wetland fraction in the tropical parts of South America and Central Africa was strongly overestimated, which seems to be related to an underestimation of potential evapotranspiration over wet tropical areas by the Penman-Monteith method used in MPI-HM. The simulated extent dynamics correlated well with monthly inundation variations obtained from satellites for most locations. Also, the simulated river discharge was affected by wetlands resulting in a delay and mitigation of peak flows. Compared to simulations without wetlands, locally increased evaporation and decreased river flow into the oceans were generated due to the implemented wetland processes.

4 Concluding remark

In the present review, we have highlighted some noticeable deficiencies in climate modelling with respect to the hydrological cycle, which provide perspectives for the modelling of relevant interactions between climate and terrestrial hydrology. These interactions are often imposed by different land-atmosphere coupling mechanisms. Over many regions, the wet state of the soil (soil moisture, wetlands, irrigation) determines feedback characteristics. These feedbacks do not only impact the local scale but also often act on the large-scale. In this respect, also human land use may affect remote regions as it has been shown for irrigation over the South Asian monsoon region. As the character of associated feedbacks vary for different regions and may change under future climate conditions, they have to be regarded in respective modelling studies. The coupling to biogeochemistry, i.e. carbon cycle and vegetation, is important to quantify feedbacks related to wetlands and permafrost. The representation of their complex dynamics within ESMs is a challenging task, but it is nevertheless necessary to investigate on-going and future climate changes over the high-latitude regions.

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References

ACIA (2005) Arctic Climate Impact Assessment. Cambridge University Press, 1042p., http://www.acia.uaf.edu.

Asharaf S, Dobler A, Ahrens B (2012) Soil moisture-precipitation feedback processes in the Indian summer monsoon season. J Hydrometeorol 13: 1461–1474, doi:10.1175/JHM-D-12-06.1.

Asokan SM, Jarsjö J, Destouni G (2010) Vapor flux by evapotranspiration: effects of changes in climate, land-use and water-use. J Geophys Res 115, D24102, doi:10.1029/2010JD014417.

Bathiany S, Claussen M, Brovkin V, Raddatz T, Gayler V (2010) Combined biogeophysical and biogeochemical effects of large-scale land-cover change in the MPI Earth system model, Biogeosciences 7: 1383-1399, doi:10.5194/bg-7-1383-2010.

Barling R, Moore I, Grayson R (1994) A quasi-dynamic wetness index for characterizing the spatial distribution of zones of surface saturation and soil water content, Water Resour Res 30: 1029–1044, doi:10.1029/93WR03346.

Bense VF, Ferguson G, Kooi H (2009) Evolution of shallow groundwater flow systems in areas of degrading permafrost. Geophys Res Letters 36, L22401, doi:10.1029/2009GL039225.Best MJ, Pryor M, Clark DB, Rooney GG, Essery RLH, Ménard CB, Edwards JM, Hendry MA, Porson A, Gedney N, Mercado LM, Sitch S, Blyth E, Boucher O, Cox PM, Grimmond CSB, Harding RJ (2011) The Joint UK Land Environment Simulator (JULES), model description – Part 1: Energy and water fluxes, Geosci Model Dev 4: 677-699, doi:10.5194/gmd-4-677-2011.

Beven K, Kirkby M (1979) A physically based, variable contributing area model of basin hydrology, Hydrol Sci Bull 24: 43–69.

Bohn T, Lettenmaier D, Sathulur K, Bowling L, Podest E, McDonald K (2007) Methane emissions from western Siberian wetlands: heterogeneity and sensitivity to climate change, Environ Res Lett 2, 045015, doi:10.1088/1748-9326/2/4/045015.

Boike J, Roth K, Overduin PP (1998) Thermal and hydrologic dynamics of the active layer at a continuous permafrost site (Taymyr Peninsula, Siberia), Water Resour Res 34(3): 355-363.

Bolton WR (2006) Dynamic modelling of the hydrologic processes in areas of discontinuous permafrost. Ph D Dissertation, 163 pp., University of Alaska, Fairbanks, US.

Bonan G (1995) Sensitivity of a GCM simulation to inclusion of inland water surfaces, J Clim 8: 2691–2704.

Boucher O, Myhre G, Myhre A (2004) Direct human influence of irrigation on atmospheric water vapor and climate. Clim Dyn 22: 597–603, doi:10.1007/s00382-004-0402-4.

Bowling L, Lettenmaier DP (2010) Modeling the Effects of Lakes and Wetlands on the Water Balance of Arctic Environments, J Hydrometeorol 11: 276–295, doi:10.1175/2009JHM1084.1.

Breckle, S-W, Geldyeva GV (2012) Dynamics of the Aral Sea in Geological and Historical Times. In S.-W. Breckle et al. (eds.), Aralkum - a Man-Made Desert: The Desiccated Floor of the Aral Sea (Central Asia), Ecological Studies 218: 13-35, DOI 10.1007/978-3-642-21117-1_2, Springer-Verlag Berlin Heidelberg.

Brown J, Ferrians Jr. OJ, Heginbottom JA, Melnikov ES (eds.) (1997) Circum-Arctic map of permafrost and ground-ice conditions. Washington, DC: U.S. Geological Survey in Cooperation with the Circum-Pacific Council for Energy and Mineral Resources. Circum-Pacific Map Series CP-45, scale 1:10,000,000.

Bullock A, Acreman M (2003) The role of wetlands in the hydrological cycle, Hydrol Earth Syst Sci 7: 358–389, doi:10.5194/hess-7-358-2003.

Cess RD, Potter GL, Zhang MH, Blanchet JP, Chalita S, Colman R, Dazlich DA, Genio AD, Dymnikov V, Galin V, Jerrett D, Keup E, Lacis AA, LE Treut H, Liang XZ, Mahfouf JF, McAvaney BJ, Meleshko VP, Mitchell JF, Morcrette JJ, Norris PM, Randall DA, Rikus L, Roeckner E, Royer JF, Schlese U, Sheinin DA, Slingo JM, Sokolov AS, Taylor KE, Washington WM, Wetherald RT, Yagai I. (1991) Interpretation of snow-climate feedback as produced by 17 general circulation models. Science 253: 888–892, doi:10.1126/science.253.5022.888.

Christensen JH (1999) Relaxation of soil variables in a regional climate model. Tellus 51A: 474–685.

Claussen M (2004) The global climate. Vegetation, Water, Humans and the Climate, P. Kabat et al., Eds., Global Change—The IGBP Series 24: 33–57, Springer.

Coe M (1997) Simulating continental surface waters: An application to holocene Northern Africa, J Clim 10: 1680–1689.

Coe M (1998) A linked global model of terrestrial hydrologic processes: simulation of modern rivers, lakes, and wetlands, J Geophys Res (D Atmos) 103: 8885–8899, doi:10.1029/98JD00347.

Coe M (2000) Modeling terrestrial hydrological systems at the continental scale: Testing the accuracy of an atmospheric GCM, J Clim 13: 686–704, doi:10.1175/1520-0442(2000)013<0686:MTHSAT>2.0.CO;2.

Coe M, Bonan G (1997) Feedbacks between climate and surface water in northern Africa during the middle Holocene, J Geophys Res (D Atmos) 102: 11087–11101, doi:10.1029/97JD00343.

Cox PM, Betts RA, Jones CD, Spall SA, Totterdell IJ (2000) Acceleration of global warming due to carbon-cycle feedbacks in a coupled climate model. Nature 408: 184-187.

Dale VH (1997) The Relationship Between Land-Use Change and Climate Change. Ecol Appl 7: 753-769.

De Ridder (2008) Soil temperature spin-up in land surface schemes. Theor Appl Climatol, doi:10.1007/s00704-008-0011-y.

De Rosnay P, Polcher J, Laval K, Sabre M (2003) Integrated parameterization of irrigation in the land surface model ORCHIDEE: Validation over Indian Peninsula, Geophys Res Lett 30: 1986, doi:10.1029/2003GL018024.

Decharme B, Douville H, Prigent C, Papa F, Aires F (2008) A new river flooding scheme for global climate applications : Offline evaluation over South America, J Geophys Res (D Atmos) 113, D11110, doi:10.1029/2007JD009376.

Decharme B, Alkama R, Papa F, Faroux S, Douville H, Prigent C (2011) Global off-line evaluation of the ISBA-TRIP flood model, Clim Dyn 38: 1–24, doi:10.1007/s00382-011-1054- 9.

Denman KL, Brasseur G, Chidthaisong A, Ciais P, Cox PM, Dickinson RE, Hauglustaine D, Heinze C, Holland E, Jacob D, Lohmann U, Ramachandran S, da Silva Dias PL, Wofsy SC, Zhang X (2007) Couplings Between Changes in the Climate System and Biogeochemistry. In: Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change [Solomon S, Qin D, Manning M, Chen Z, Marquis M, Averyt K.B, Tignor M, Miller HL (eds.)]. Cambridge University Press, Cambridge, United Kingdom and New York, NY, USA.

Destouni G, Jaramillo F, Prieto C (2013) Hydroclimatic shifts driven by human water use for food and energy production. Nature Clim Change 3: 213-217.

Dirmeyer P; Koster R, Guo ZAD (2006) Do Global Models Properly Represent the Feedback between Land and Atmosphere? J Hydrometeorol 7: 1177–1198.

Döll P, Hoffmann-Dobrev H, Portmann FT, Siebert S, Eicker A, Rodell M, Strassberg G, Scanlon BR (2012) Impact of water withdrawals from groundwater and surface water on continental water storage variations. J Geodyn 59/60: 143-156.

Douglas EM, Niyogi D, Frolking S, Yeluripati JB, Pielke RA Sr, Niyogi N, Vörösmarty CJ, Mohanty UC (2006) Changes in moisture and energy fluxes due to agricultural land use and irrigation in the Indian Monsoon Belt. Geophys Res Lett 33: L14403. doi:10.1029/2006GL026550.

Douglas EM, Beltrán-Przekurat A, Niyogi D, Pielke RA Sr., Vörösmarty CJ (2009) The impact of agricultural intensification and irrigation on land–atmosphere interactions and Indian monsoon precipitation—A mesoscale modeling perspective, Global Planet Change 67: 117–128, doi:10.1016/j.gloplacha.2008.12.007.

Ekici A, Beer C, Hauck C, Hagemann S (2013) Improved soil physics for simulating high latitude permafrost regions by the JSBACH terrestrial ecosystem model. Geosci Model Dev Discuss 6: 2655-2698, doi:10.5194/gmdd-6-2655-2013.

Falloon P, Betts RA (2010) Climate impacts on European agriculture and water management in the context of adaptation and mitigation—The importance of an integrated approach. Sci Total Environ 408: 5667–5687.

Feddema JJ, Oleson KW, Bonan GB, Mearns LO, Buja LE, Meehl GA, Washington WM (2005) The importance of land-cover change in simulating future climates. Science 310: 1674-1678.

Fischer EM, Seneviratne SI, Vidale PL, Lüthi D, Schär C (2007) Soil moisture - atmosphere interactions during the 2003 European summer heat wave. J Clim 20: 5081-5099.

Frampton A, Painter SL, Lyon SW, Destouni G (2011) Nonisothermal, three-phase simulations of near-surface flows in a model permafrost system under seasonal variability and climate change. J Hydrol 403: 352–359, doi:10.1016/j.jhydrol.2011.04.010.

Frampton A, Painter SL, Destouni G (2013) Permafrost degradation and subsurface-flow changes caused by surface warming trends, Hydrogeol J 21: 271-280.

French HM (1990) Editorial. Permafrost Periglac Process 1: 1, doi: 10.1002/ppp.3430010102.

French HM (2007) The periglacial environment - 3rd edition. John Wiley & Sons Ltd., West Sussex, UK, 457 pp.

Friborg T, Soegaard H, Christensen T, Lloyd C, Panikov N (2003) Siberian wetlands: Where a sink is a source, Geophys Res Lett 30: CLM 5–1–CLM 5–4, doi:10.1029/2003GL017797.

Gao X, Luo Y, Lin W, Zhao Z, Giorgi F (2003) Simulation of effects of land use change on climate in China by a regional climate model. Adv Atmos Sci 20 (4): 583-592.

Gedney N, Cox P, Huntingford C (2004) Climate feedback from wetland methane emissions. Geophys Res Lett 31, L20503, doi:10.1029/2004GL020919.

Gerten D. (2013) A vital link: water and vegetation in the Anthropocene. Hydrol Earth Syst Sci Discuss 10: 4439–4462, doi:10.5194/hessd-10-4439-2013.

Gerten D, Rost S, von Bloh W, Lucht W (2008): Causes of change in 20th century global river discharge. Geophys Res Lett 35, L20405, doi:10.1029/2008GL035258.

Gerten D, Hagemann S, Biemans H, Saeed F, Konzmann M (2011) Climate change and irrigation: feedbacks and impacts. WATCH Tech. Rep. 47.

Göttel H, Alexander J, Keup-Thiel E, Rechid D, Hagemann S, Blome T, Wolf A, Jacob D (2008) Influence of changed vegetations fields on regional climate simulations in the Barents Sea Region. Clim Change 87: 35-50, doi: 10.1007/s10584-007-9341-5.

Gordon LJ, Steffen W, Jonsson BF, Folke C, Falkenmark M, Johannessen A (2005): Human modification of global water vapour flows from the land surface. P Natl Acad Sci USA 102: 7612–7617.

Gorham E (1991) Northern peatlands: role in the carbon cycle and probable responses to climatic warming, Ecol Appl 1: 182–195, doi:10.2307/1941811.

Gouttevin I, Krinner G, Ciais P, Polcher J, Legout C (2012) Multi-scale validation of a new soil freezing scheme for a land-surface model with physically-based hydrology. The Cryosphere 6: 407-430, doi:10.5194/tc-6-407-2012.

Guimberteau M, Laval K, Perrier A, Polcher J (2012) Global effect of irrigation and its impact on the onset of the Indian summer monsoon. Clim Dyn 39: 1329-1348.

Harding RJ, Best M, Blyth E, Hagemann S, Kabat P, Tallaksen LM, Warnaars T, Wiberg D, Weedon GP, van Lanen H, Ludwig F, Haddeland I (2011) Current knowledge of the terrestrial Global Water Cycle. J Hydrometeorol 12: 1149–1156, doi: 10.1175/JHM-D-11-024.1.

Hashimoto S, Wattenbach M, Smith P (2011) A new scheme for initializing process-based ecosystem models by scaling soil carbon pools. Ecolog Modelling 222: 3598– 3602.

Haugen RK, Slaughter CW, Howe KE, Dingman SL (1982) Hydrology and climatology of the Caribou-Poker Creeks Research Watershed, Alaska, CRREL Rep. 82-26, 42 pp., 1982.

Hinzman LD, Kane DI (1991) Snow hydrology of a headwater arctic basin, 2. Conceptual analysis and computer modeling, Water Resour Res 27(6): 95-100.

International Permafrost Association (1998) Circumpolar active-layer permafrost system (CAPS), version 1.0.

Jacob D, Bärring L, Christensen OB, Christensen JH, Hagemann S, Hirschi M, Kjellström E, Lenderink G, Rockel B, Schär C, Seneviratne SI, Somot S, van Ulden A, van den Hurk B (2007) An inter-comparison of regional climate models for Europe: Design of the experiments and model performance. Clim Change 81 (Suppl 1): 31-52.

Karim A, Veizer J (2002) Water balance of the Indus River Basin and moisture source in the Karakoram and western Himalaya: Implications from hydrogen and oxygen isotopes in river water, J Geophys Res 107(D18): 4362, doi:10.1029/2000JD000253.

Khan S, Rana T, Gabriel HF, Ullah M (2008) Hydrogeologic assessment of escalating groundwater exploitation in the Indus Basin, Pakistan, Hydrogeol J 16: 1635–1654, doi:10.1007/s10040-008-0336-8.

Kleinen T, Brovkin V, Getzieh R (2012) A dynamic model of wetland extent and peat accumulation: results for the Holocene, Biogeosciences 9: 235–248, doi:10.5194/bg-9-235-2012.

Koren V, Schaake J, Mitchell K, Duan OY, Chen F, Baker JM (1999) A parameterization of snowpack and frozen ground intended for NCEP weather and climate models. J Geophys Res 104: 19569–19585.

Koster RD, Dirmeyer PA, Guo Z, Bonan G, Chan E, Cox P, Gordon CT, Kanae S, Kowalczyk E, Lawrence D, Liu P, Lu CH, Malyshev S, McAvaney B, Mitchell K, Mocko D, Oki T, Oleson K, Pitman A, Sud YC, Taylor CM, Verseghy D, Vasic R, Xue Y, Yamada T (2004) Regions of strong coupling between soil moisture and precipitation. Science 305: 1138–1140.

Koster RD, Guo Z, Dirmeyer PA, Bonan G, Chan E, Cox P, Davies H, Gordon CT, Kanae S, Kowalczyk E, Lawrence D, Liu P, Lu CH, Malyshev S, McAvaney B, Mitchell K, Mocko D, Oki T, Oleson KW, Pitman A, Sud YC, Taylor CM, Verseghy D, Vasic R, Xue Y, Yamada T (2006) GLACE: The Global Land-Atmosphere Coupling Experiment. Part I: Overview, J Hydrometeorol 7: 590–610.

Koven CD, Riley WJ, Stern A (2012) Analysis of permafrost thermal dynamics and response to climate change in the CMIP5 Earth System Models. J Clim, doi:10.1175/JCLI-D-12-00228.1, in press.

Krinner G, Lézine A-M, Braconnot P, Sepulchre P, Ramstein G, Grenier C, Gouttevin I (2012) A reassessment of lake and wetland feedbacks on the North African Holocene climate, Geophys Res Lett 39, L07701, doi:10.1029/2012GL050992.

Kumar P, Wiltshire A, Mathison C, Asharaf S, Ahrens B, Lucas-Picher P, Christensen JH, Gobiet A, Saeed F, Hagemann S, Jacob D (2013) Downscaled climate change projections with uncertainty assessment over India using a high resolution multi-model approach. Sci Total Environ (2013), http://dx.doi.org/10.1016/j.scitotenv.2013.01.051.

Lawrence DM, Oleson KW, Flanner MG, Thornton PE, Swenson SC, Lawrence PJ, Zeng X, Yang Z-L, Levis S, Sakaguchi K, Bonan GB, Slater AG (2011) Parameterization improvements and functional and structural advances in version 4 of the community land model. J Adv Model Earth Syst 3, 2011MS000 045, doi:10.1029/2011MS000045.

Lee E, Chase T, Rajagopalan B, Barry R, Wiggs T, Lawrence P (2009) Effects of irrigation and vegetation activity on early Indian summer monsoon variability, Int J Climatol 29: 573–581, doi:10.1002/joc.1721.

Levin D, Hopper I, Ernst F, Trivett M, Worthy N (2000) Evidence for a link between climate and northern wetland methane emissions, J Geophys Res (D Atmos) 105: 4031–4038, doi:10.1029/1999JD901100.

Lobell D, Bala G, Mirin A, Phillips T, Maxwell R, Rotman D (2009) Regional differences in the influence of irrigation on climate. J Clim 22: 2248–2255, doi:10.1175/2008JCLI2703.1.

Loew A, Holmes T, de Jeu R (2009) The European heat wave 2003: early indicators from multisensoral microwave remote sensing? J Geophys Res 114: D05103.

Luo LF, Robock A, Vinnikov KY, Schlosser CA, Slater AG, Boone A, Braden H, Cox P, de Rosnay P, Dickinson RE, Dai YJ, Duan QY, Etchevers P, Henderson-Sellers A, Gedney N, Gusev YM, Habets F, Kim JW, Kowalczyk E, Mitchell K, Nasonova ON, Noilhan J, Pitman AJ, Schaake J, Shmakin AB, Smirnova TG, Wetzel P, Xue YK, Yang ZL, Zeng QC (2003) Effects of frozen soil on soil temperature, spring infiltration, and runoff: Results from the PILPS 2(d) experiment at Valdai, Russia. J Hydrometeorol 4: 334–351.

Lucas-Picher P, Christensen JH, Saeed F, Kumar P, Asharaf S, Ahrens B, Wiltshire A, Jacob D, Hagemann S (2011) Can regional climate models represent the Indian monsoon? J Hydrometeorol 12: 849-868, doi:10.1175/2011JHM1327.1.

Mackay JR (1988) Catastrophic lake drainage, Tuktoyaktuk peninsula area, District of Mackenzie. Geol Surv Canada, paper 88-1D: 83-90.

Mahajan PN, Talwalkar DR, Chinthalu GR, Rajamaniet S (1995) Use of INSAT winds for better depiction of monsoon depression over Indian region. Meteorol Appl 2: 333–339.

Marengo J, Nobre CA (2001) The hydroclimatological framework in Amazonia. In: Biogeochemistry of the Amazon Basin [McClaine M, Victoria R, Richey J (eds.)]. Oxford University Press, Oxford, UK, 17–42.

McGuire AD, Anderson LG, Christensen TR, Dallimore S, Guo L, Hayes DJ, Heimann M, Lorenson TD, Macdonald RW, Roulet N (2009) Sensitivity of the carbon cycle in the Arctic to climate change. Ecological Monographs 79: 523–555, doi:10.1890/08-2025.1.

Meehl GA, Boer GJ, Covey C, Latif M, Stouffer RJ (2000) The Coupled Model Intercomparison Project (CMIP). Bull Amer Meteor Soc 81: 313–318.

Mishra V, Cherkauer K, Bowling L (2010) Parameterization of lakes and wetlands for energy and water balance studies in the great lakes region, J Hydrometeorol 11: 1057–1082, doi:10.1175/2010JHM1207.1.

Niu G-Y, Yang Z-L (2006) Effects of frozen soil on snowmelt runoff and soil water storage at a continental scale. J Hydrometeorol 7: 937–952, doi:10.1175/JHM538.1.

O'Connor F, Boucher O, Gedney N, Jones C, Folberth G, Coppell R, Friedlingstein P, Collins W, Chappellaz J, Ridley J, Johnson C (2010) Possible role of wetlands, permafrost, and methane hydrates in the methane cycle under future climate change: A review, Rev Geophys 48, RG4005, doi:10.1029/2010RG000326.

Oleson K, Dai Y, Bonan GB, Bosilovich M, Dickinson R, Dirmeyer P, Hoffman F, Houser P, Levis S, Niu G-Y, Thornton P, Vertenstein M, Yang Z-L, Zeng X (2004): Technical Description of the Community Land Model (CLM). NCAR Technical Note NCAR/TN-461+STR, doi:10.5065/D6N877R0.

Oyama MD, Nobre CA (2004) Climatic consequences of a largescale desertifi cation in northeast Brazil: a GCM simulation study. J. Clim. 17: 3203–3213.

Paeth H, Born K, Girmes R, Podzun R, Jacob D (2009) Regional climate change in tropical and northern Africa due to greenhouse forcing and land-use changes. J Clim 22: 114–132, doi:10.1175/2008JCLI2390.1.

Piao S, Friedlingstein P, Ciais P, de Noblet-Ducoudré N, Labat D, Zaehle S (2007): Changes in climate and land use have a larger direct impact than rising CO2 on global river runoff trends. P Natl Acad Sci USA 104: 15242–15247.

Pielke Sr. RA, Marland G, Betts RA, Chase TN, Eastman JL, Niles JO, Niyogi DDS, Running SW (2002) The influence of land-use change and landscape dynamics on the climate system: relevance to climate-change policy beyond the radiative effect of greenhouse gases. Phil Trans R Soc Lond A 360: 1705-1719, doi:10.1098/rsta.2002.1027.

Puma MJ, Cook BI (2010) Effects of irrigation on global climate during the 20th century. J Geophys Res 115, D16120, doi:10.1029/2010JD014122.

Richards LA (1931) Capillary conduction of liquids through porous mediums. Physics 1 (5): 318–333: doi:10.1063/1.1745010.

Ringeval B, De Noblet-Ducoudré N, Ciais P, Bousquet P, Prigent C, Papa F, Rossow W (2010) An attempt to quantify the impact of changes in wetland extent on methane emissions on the seasonal and interannual time scales, Global Biogeochem Cycles 24, GB2003, doi:10.1029/2008GB003354.

Ringeval B, Friedlingstein P, Koven C, Ciais P, de Noblet- Ducoudré N, Decharme B, Cadule P (2011) Climate-CH4 feedback from wetlands and its interaction with the climate-CO2 feedback, Biogeosciences 8: 2137–2157, doi:10.5194/bg- 8-2137-2011.

Riseborough D, Shiklomanov N, Etzelmüller B, Gruber S, Marchenko S (2008) Recent advances in permafrost modelling. Permafr Periglac Process 19: 137–156, doi:10.1002/ppp.615.

Romanovskii NN, Hubberten H-W, Gavrilov AV, Tumskoy VE, Tipenko GS, Grigoriev MN (2000) Thermokarst and land-ocean interactions, Laptev Sea region, Russia. Permafrost Periglacial Processes 11: 137-152.

Rost S, Gerten D, Bondeau A, Lucht W, Rohwer J, Schaphoff S (2008a): Agricultural green and blue water consumption and its influence on the global water system. Water Resour Res 44, W09405, doi:10.1029/2007WR006331.

Rost S, Gerten D, Heyder U (2008b): Human alterations of the terrestrial water cycle through land management, Adv Geosci 18: 43–50, doi:10.5194/adgeo-18-43-2008.

Sacks WJ, Cook BI, Buenning N, Levis S, Helkowski JH (2009) Effects of global irrigation on the near‐surface climate, Clim Dyn 33: 159–175, doi:10.1007/s00382-008-0445-z.

Saeed F, Hagemann S, Jacob D (2009) Impact of irrigation on the South Asian Summer Monsoon. Geophys Res Letters 36, L20711, doi:10.1029/2009GL040625.

Saeed F, Hagemann S, Jacob D (2011) A framework for the evaluation of the South Asian Summer Monsoon in a regional climate model applied to REMO. Int J Climatol, doi:10.1002/joc.2285.

Saeed F, Hagemann S, Saeed S, Jacob D (2012) Influence of mid-latitude circulation on upper Indus basin precipitation: the explicit role of irrigation. Clim Dyn, doi:10.1007/s00382-012-1480-3.

Schramm I, Boike J, Bolton WR, Hinzman LD (2007) Application of TopoFlow, a spatially distributed hydrological model, to the Imnavait Creek watershed, Alaska. J Geophys Res 112, G04S46, doi:10.1029/2006JG000326.

Schuur EAG, Bockheim J, Canadell JG, Euskirchen E, Field CB, Goryachkin SV, Hagemann S, Kuhry P, Lafleur PM, Lee H, Mazhitova G, Nelson FE, Rinke A, Romanovsky VE, Shiklomanov N, Tarnocai C, Venevsky S, Vogel JG, Zimov SA (2008) Vulnerability of permafrost carbon to climate change: Implications for the global carbon cycle. Bioscience 58 : 701-714, doi:10.1641/B580807.

Seneviratne SI, Stöckli R (2008) The role of land-atmosphere interactions for climate variability in Europe. In: Climate Variability and Extremes during the Past 100 years, Brönnimann et al. (eds.), Adv Global Change Res 33, Springer Verlag. (Book chapter)

Seneviratne SI, Lüthi D, Litschi M, Schär C (2006) Land-atmosphere coupling and climate change in Europe. Nature 443: 205-209.

Seneviratne SI, Corti T, Davin E, Hirschi M, Jaeger EB, Lehner I, Orlowsky B, Teuling AJ

(2010) Investigating soil moisture-climate interactions in a changing climate: A review. Earth-Sci Rev 99: 125-161, doi:10.1016/j.earscirev.2010.02.004.

Serreze MC, Barry RG (2011) Processes and impacts of Arctic amplification: A research synthesis. Global Planet Change 77: 85-96, doi:10.1016/j.gloplacha.2011.03.004.

Siebert S, Döll P, Hoogeveen J, Faurés J-M, Frenken K, Feick S (2005) Development and validation of the global map of irrigation areas. Hydrol Earth Syst Sci 9:535–547.

Singh D, Bhadram CVV, Mandal GS (1995) New regression model for Indian summer monsoon rainfall, Meteorol Atmos Phys 55: 77– 86, doi:10.1007/BF01029603.

Sitch S, Smith B, Prentice IC, Arneth A, Bondeau A, Cramer W, Kaplan JO, Levis S, Lucht W, Sykes MT, Thonicke K, Venevsky S (2003) Evaluation of ecosystem dynamics, plant geography and terrestrial carbon cycling in the LPJ Dynamic Global Vegetation Model. Glob Change Biol 9: 161–185.

Smith LC, Sheng Y, MacDonald GM, Hinzman LD (2005) Disappearing Arctic lakes. Science 308: 1429.

Solomon S, Qin D, Manning M, Marquis M, Averyt K, Tignor MMB, Miller Jr. HL, Chen Z (Eds.) (2007) Climate change 2007: The physical science basis, Cambridge University Press, 996 pp.

St-Hilaire F, Wu J, Roulet N, Frolking S, Lafleur P, Humphreys E, Arora V (2010) McGill wetland model: Evaluation of a peatland carbon simulator developed for global assessments, Biogeosciences 7: 3517–3530, doi:10.5194/bg-7-3517-2010.

Stacke T, Hagemann S (2012) Development and validation of a global dynamical wetlands extent scheme. Hydrol Earth Syst Sci 16: 2915-2933, doi:10.5194/hess-16-2915-2012.

Staehli M, Jansson P-E, Lundin LC (1999) Soil moisture redistribution and infiltration in frozen sandy soils. Water Resour Res 35: 95–103. doi: 10.1029/1998WR900045.

Swenson SC, Lawrence DM, Lee H (2012) Improved simulation of the terrestrial hydrological cycle in permafrost regions by the Community Land Model. J Adv in Modelling Earth Systems 4, doi:10.1029/2012MS000165.

Tarnocai C, Canadell JG, Schuur EAG, Kuhry P, Mazhitova G, Zimov S (2009) Soil organic carbon pools in the northern circumpolar permafrost region. Global Biogeochem Cycles 23, doi:10.1029/2008GB003327.

Taylor CM, Lambin EF, Stephenne N, Harding RJ, Essery RLH (2002) The influence of land use change on climate in the Sahel. J Clim 15: 3615–3629.

Ter Maat HW, Hutjes RWA, Ohba R, Ueda H, Bisselink B, Bauer T (2006) Meteorological impact assessment of possible large scale irrigation in Southwest Saudi Arabia. Global Planet Change 54: 183-201.

Teuling AJ, Hirschi M, Ohmura A, Wild M, Reichstein M, Ciais P, Buchmann N, Ammann C, Montagnani L, Richardson AD, Wohlfahrt G, Seneviratne SI (2009) A regional perspective on trends in continental evaporation. Geophys Res Lett 36, L02404, doi:10.1029/2008GL036584.

Tuinenburg OA, Hutjes RWA, Stacke T, Wiltshire A, Lucas-Picher P (2013) Effects of Irrigation in India on the Atmospheric Water Budget. J Hydrometeorol, submitted.

Uppala SM, Kallberg PW, Simmons AJ, Andrae U, Da Costa Bechtold V, Fiorino M, Gibson JK, Haseler J, Hernandez A, Kelly GA, Li X, Onogi K, Saarinen ACM, Sokka N, Allan RP, Andersson E, Arpe K, Balmaseda MA, BeljaarsACM, Van de Berg L, Bidlot J, Bormann N, Caires S, Dethof A, Dragosavac M, Fisher M, Fuentes M, Hagemann S, Holm E, Hoskins BJ, Isaksen L, Janssen PAEM, McNally AP, Mahfouf JF, Jenne R, Morcrette JJ, Rayner NA, Saunders RW, Simon P, Sterl A, Trenberth KE, Untch A, Vasiljevic D, Viterbo P, Woollen J (2005) The Era-40 Re-analysis. Q J R Meteorol Soc 131: 2961–3012, doi:10.1256/qj.04.176.

van der Velde Y, Lyon SW, Destouni G (2013) Data-driven regionalization of river discharges and emergent land cover-evapotranspiration relationships across Sweden. J Geophys Res 118: 1–12, doi:10.1002/jgrd.50224.

Verseghy DL (1991) Class - a Canadian land surface scheme for GCMs. i. soil model. Int J Climatol 11: 111-133.

Vörösmarty C, Gutowski W, Person M, Chen T-C, Case D (1993) Linked atmosphere-hydrology models at the macroscale, in: Wilkinson W (ed.) Macroscale Modeling of the Hydrosphere, no. 214 in International Association of Hydrological Sciences: 3-27.

Walker HJ (1978) Lake tapping in the Colville River delta. In: Permafrost, Proceedings of the Third International Confrerence on Permafrost, 10-13 July, Edmonton, Alberta, Canada. National Research Council of Canada, Ottawa, vol. 1: 233-238.

Walsh JE, Anisimov O, Hagen JOM, Jakobsson T, Oerlemans J, Prowse T.D, Romanovsky V, Savelieva N, Serreze M, Shiklomanov A, Shiklomanov I, Solomon S, Arendt A, Atkinson D, Demuth MN, Dowdeswell J, Dyurgerov M, Glazovsky A, Koerner RM, Meier M, Reeh N, Sigurosson O, Steffen K, Truffer M (2005) Cryosphere and hydrology, in: Symon C, Arris L, Heal B (eds.) Arctic Climate Impact Assessment, Chap. 6: 184-242, Cambridge University Press.

Willmott CJ, Matsuura K (2009) Terrestrial Precipitation: 1900–2008 Gridded Monthly Time Series (1900– 2008), http://climate.geog.udel.edu, Univ. of Del., Newark.

Woo MK (1986) Permafrost hydrology in North America, Atmosphere-Ocean 24(3): 201-234.

Woo M, Kane DL, Carey SK, Yang D (2008) Progress in permafrost hydrology in the new millennium. Permafr Periglac Process 19: 237–254.

Wutzler T, Reichstein M (2007) Soils apart from equilibrium – consequences for soil carbon balance modelling. Biogeosci 4: 125–136.

Yoshikawa K, Bolton WR, Romanovsky VE, Fukuda M, Hinzman LD (2002), Impacts of wildfire on the permafrost in the boreal forest of interior Alaska, J Geophys Res 107, 8148, doi:10.1029/2001JD000438, [printed 108 (D1), 2003].

Yu Z, Pollard D, Cheng L (2006) On continental-scale hydrologic simulations with a coupled hydrologic model, J Hydrol 331: 110–124, doi:10.1016/j.jhydrol.2006.05.021.

Zimov SA, Davydov SP, Zimova GM, Davydova AI, Schuur EAG, Dutta K, Chapin III FS (2006) Permafrost carbon: Stock and decomposability of a globally significant carbon pool. Geophys Res Lett 33, doi:10.1029/2006GL027484.

Figure captions

Fig. 1 **a)** Fraction of area equipped for irrigation in South Asia based on the reference year 2000 (Siebert et al. 2005). **b)** Locations of Indus and Ganges catchments at 0.5° resolution.

Fig. 2 Simulated (upper panels) and observed (lower panels) summer climatologies (JJAS) for the period 1961-2000: 2m Temperature [°C] (left, REMO and data of Wilmott and Matsuura 2009) and mean sea level pressure [hPa] (right, REMO and ERA40. The white circles indicate the heat low region.

Fig. 3 Simulated REMO summer climatologies (JJAS) for the period 1989-1992: 2m Temperature [°C], **a)** Reference, **b)** REMO with irrigation; and mean sea level pressure [hPa], **c)** Reference, **d)** REMO with irrigation.

Fig. 4 Differences of several variables from the REMO simulations with and without irrigation averaged for the summer (JJAS) 1989–1992: **a)** Mean Sea Level Pressure (hPa), **b)** 2m temperature (°C), **c)** Evapotranspiration (mm/day), and **d)** 850 hPa winds. The shaded blue areas indicate significant differences at the 90% level from a two tailed t-test.

Fig. 5 Simulated development and movement of a monsoon depression that has been observed within the period from 18-22 September 1991: The panels show the MSLP contour lines at 999 hPa around the centre of the depression as simulated by REMO without (left) and with (right) irrigation .

Fig. 6 Projected 2m temperature changes (2085-2099 minus 1985-1999) in °C for REMO without (left panel) and with irrigation (right panel).

Fig. 7 Distribution of permafrost areas in the Arctic according to the International Permafrost Association (1998).

Fig. 8 Simulated total permafrost area for historical 20th century climate and future climate following the RCP8.5 scenario for various CMIP5 models. Figure is taken from Koven et al. (2012).

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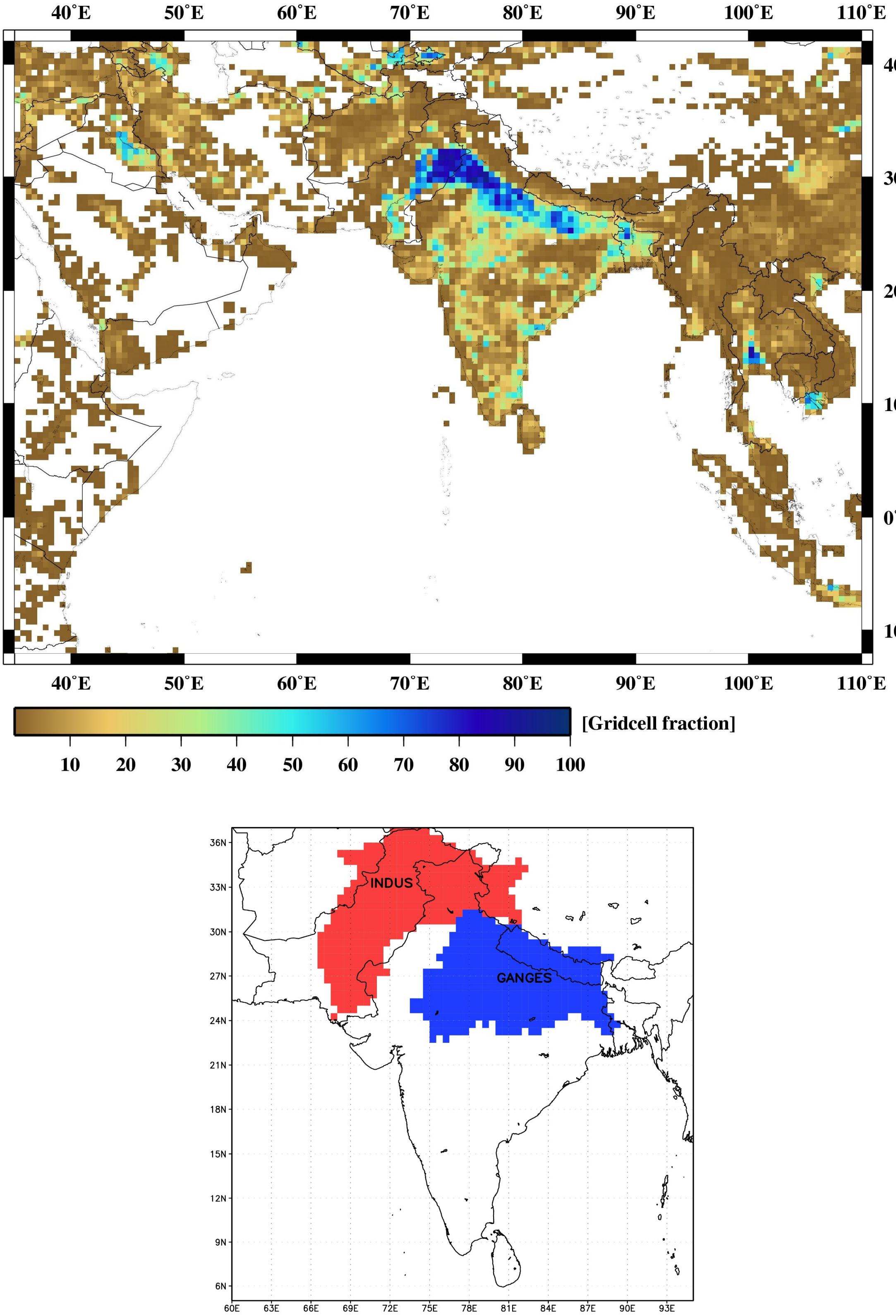
a)

b)



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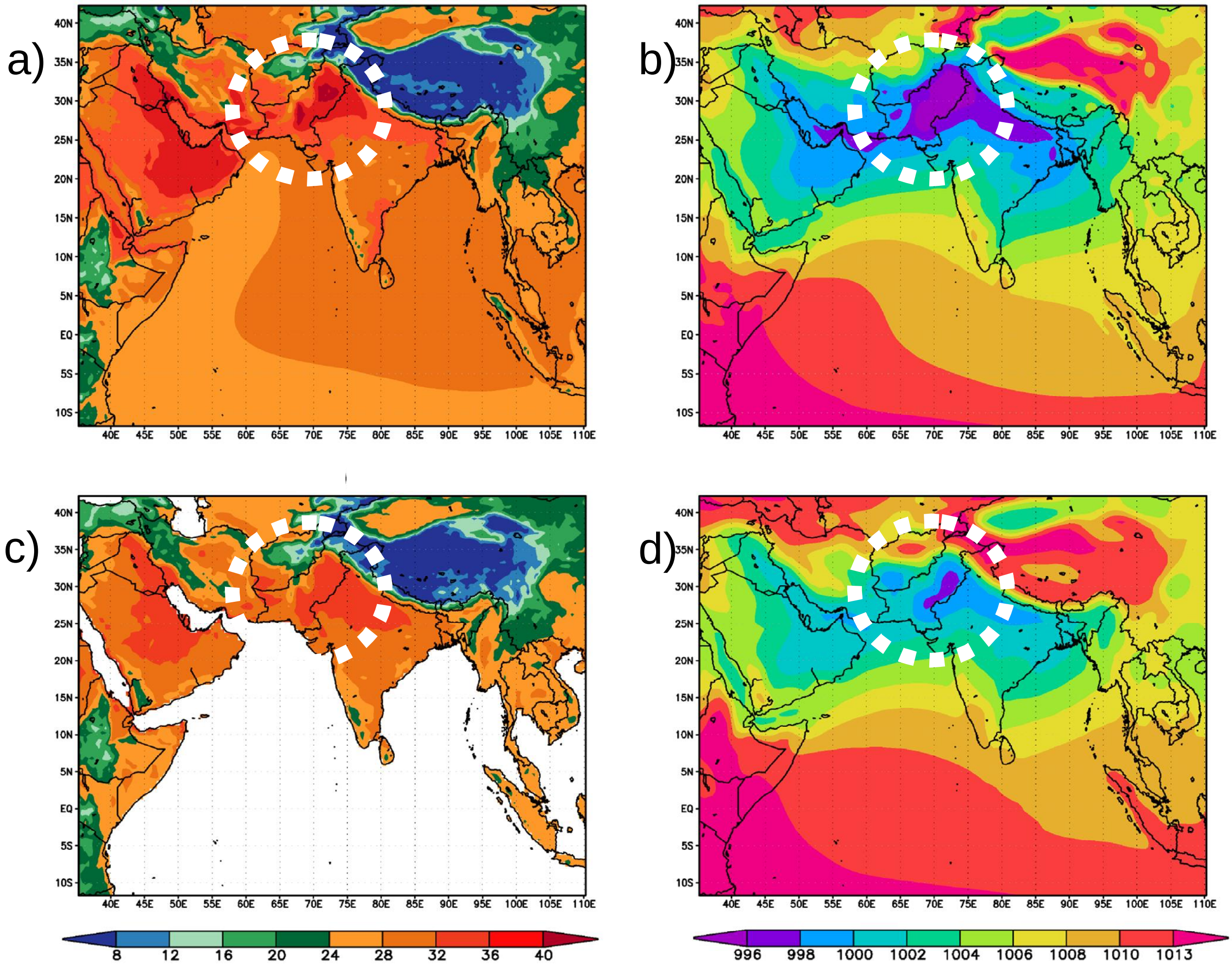

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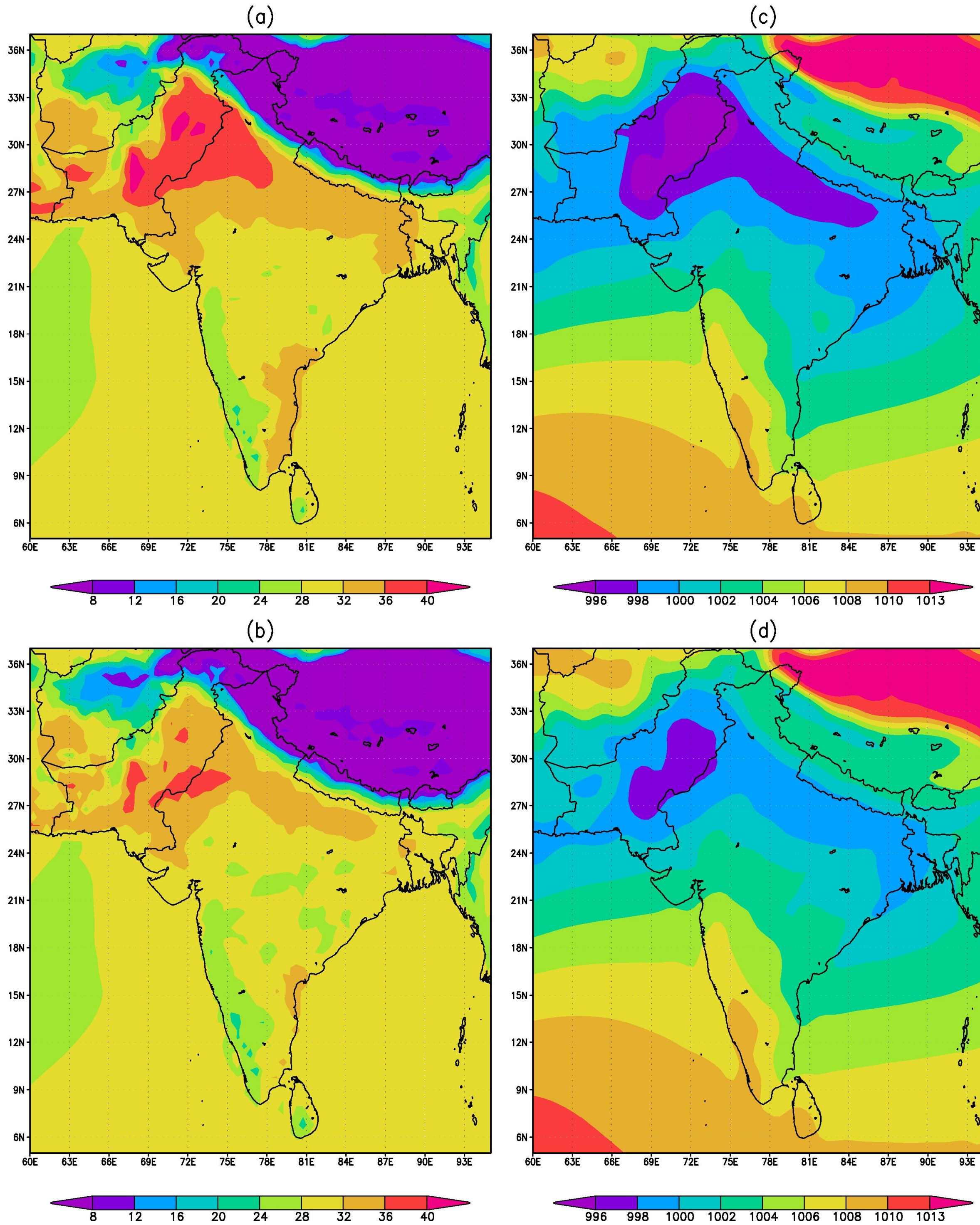

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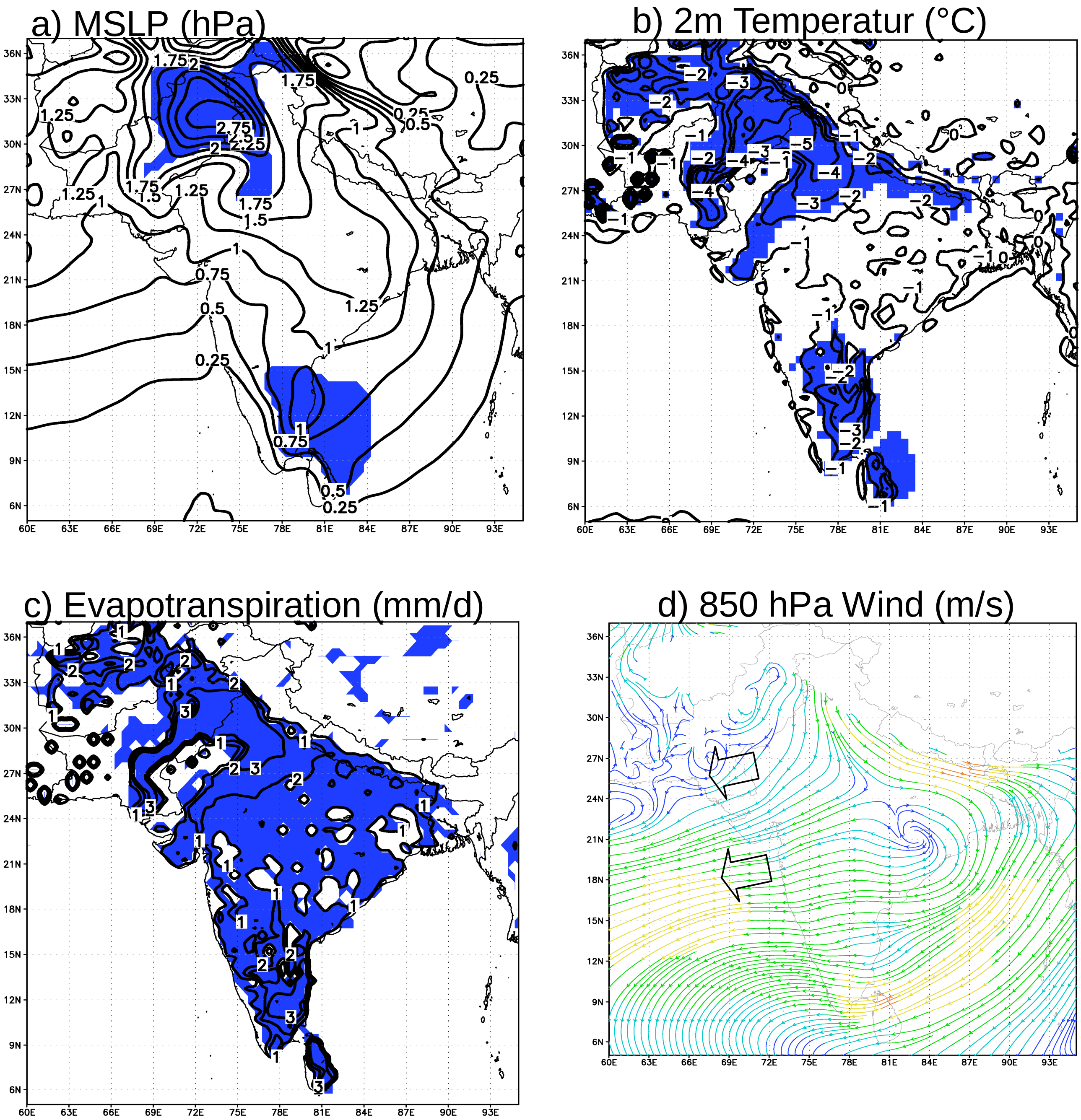

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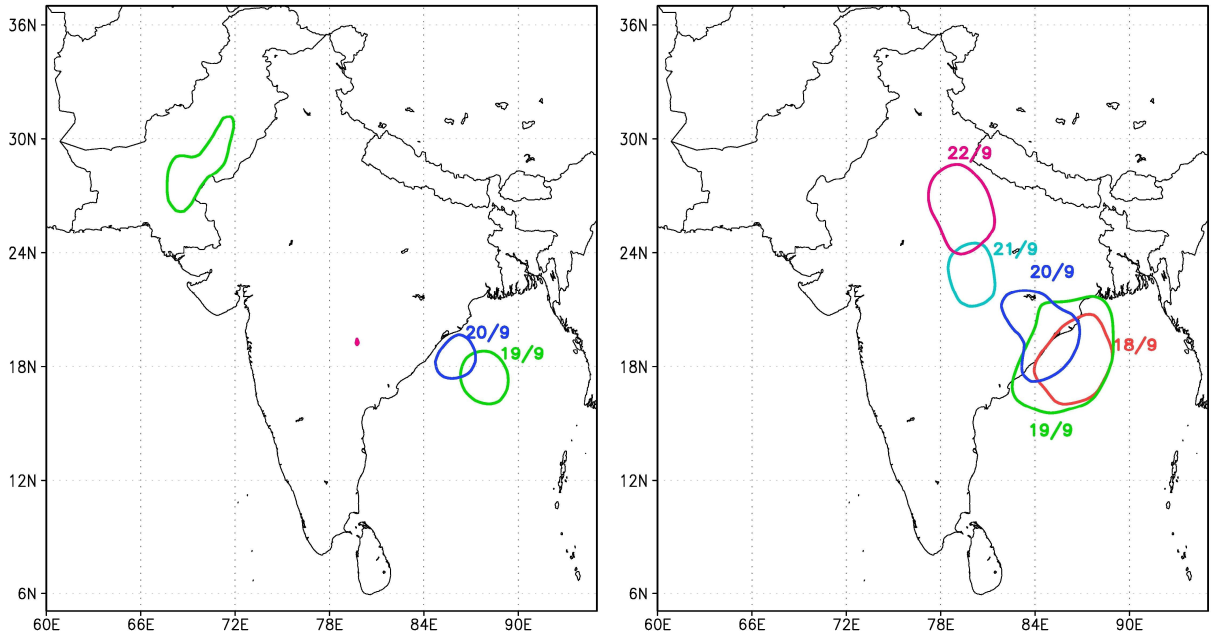

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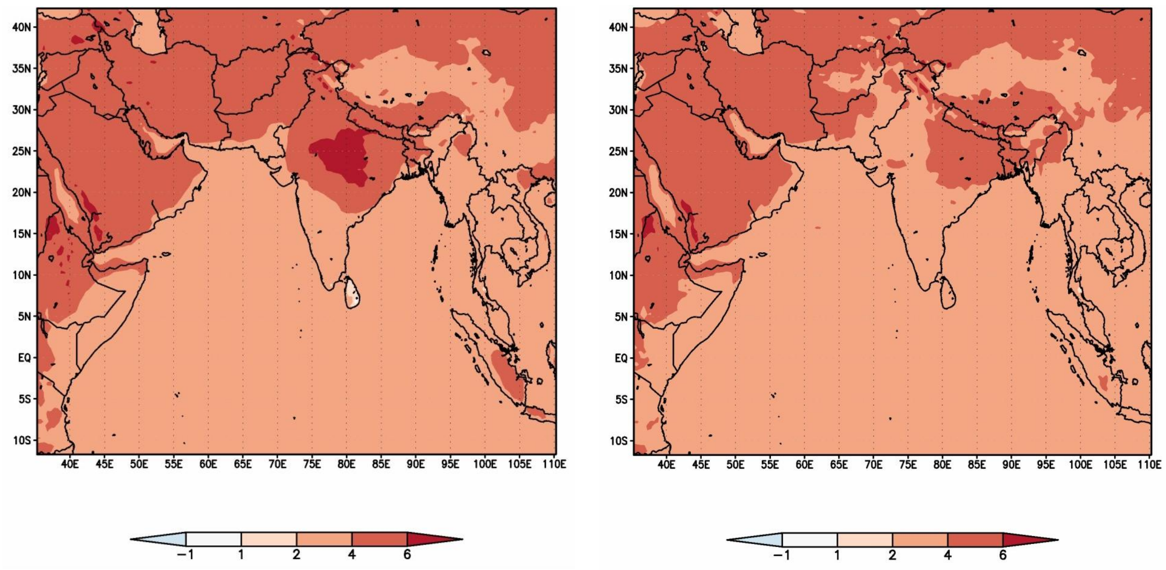

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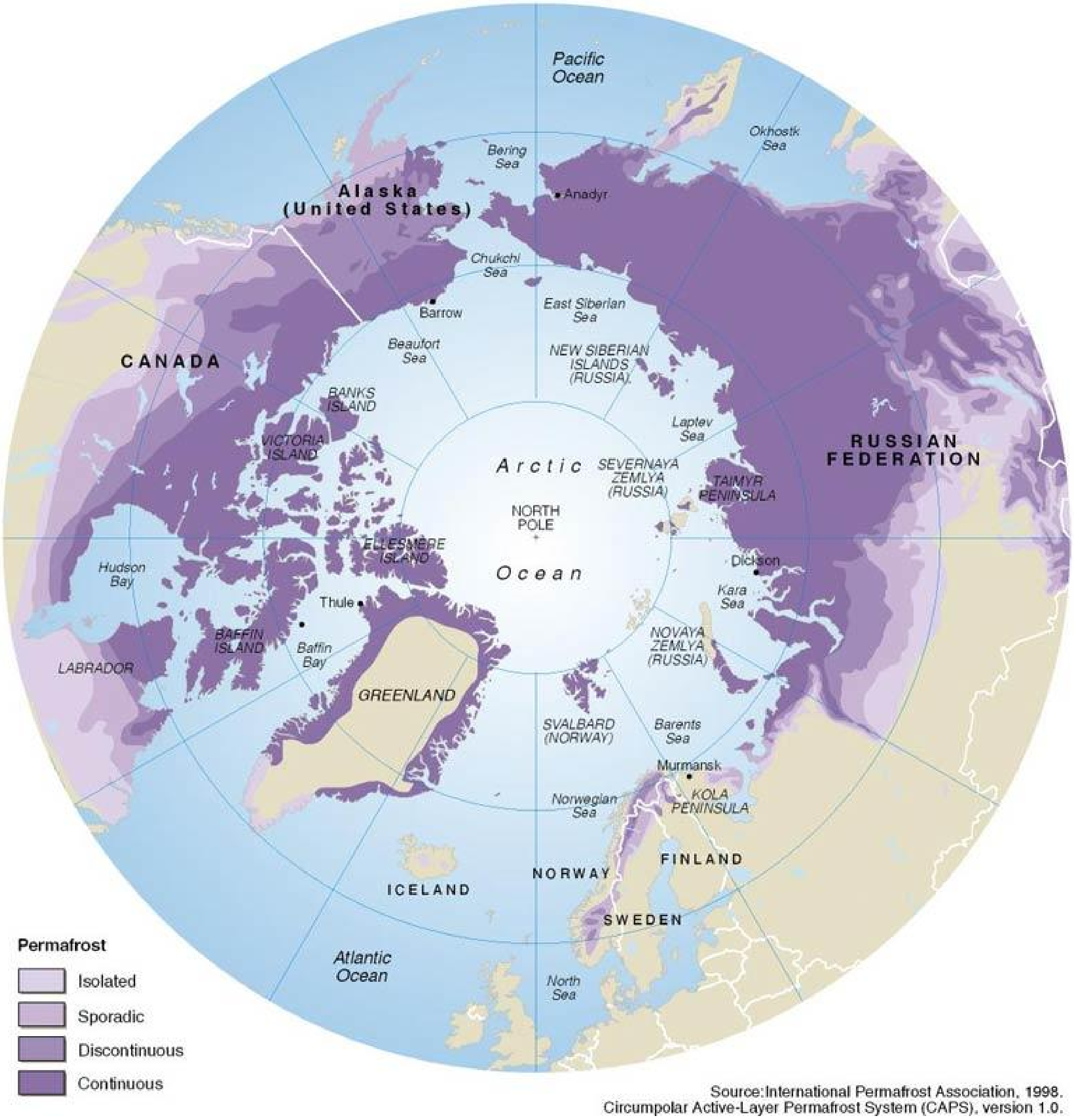

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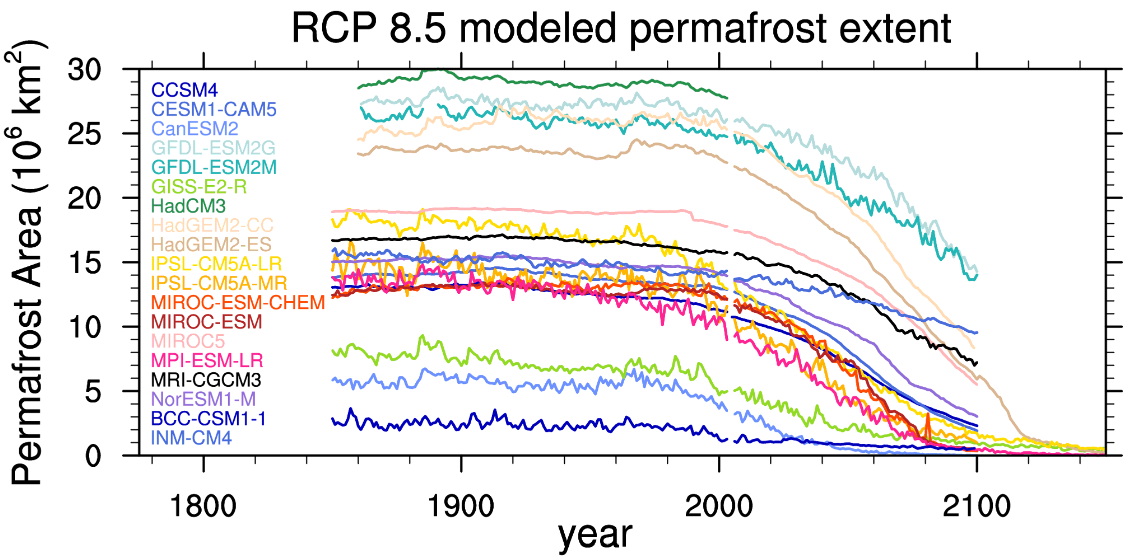

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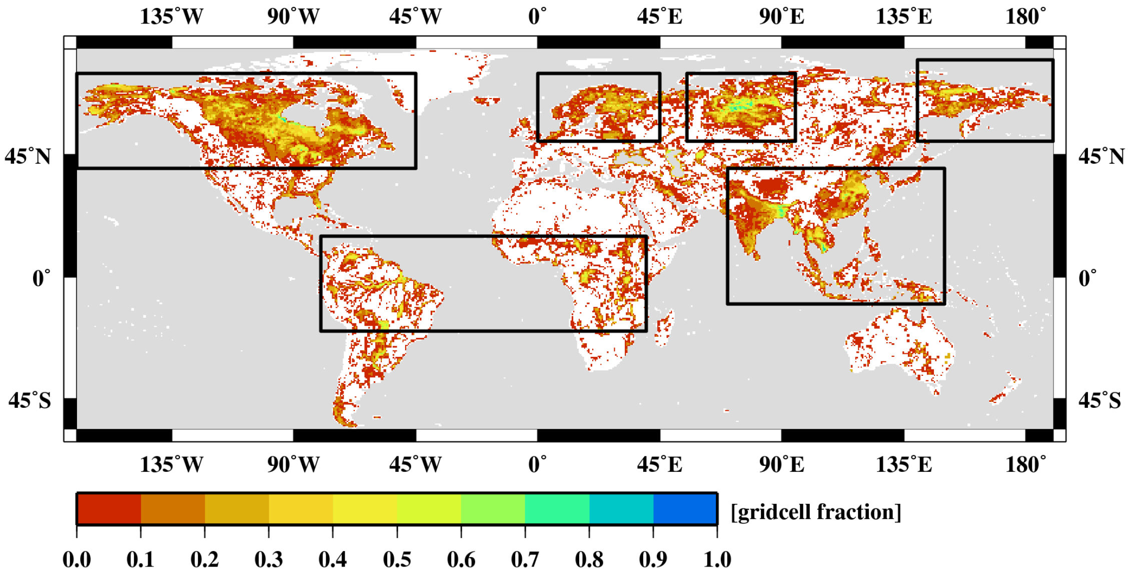

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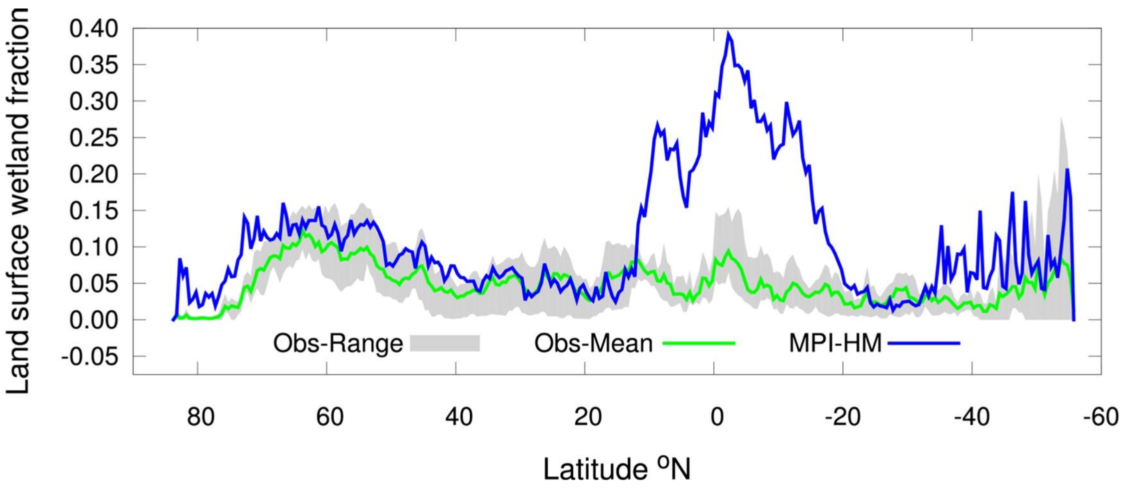

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