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Spatial and temporal connections in groundwater contribution to evaporation

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Abstract

In climate models, lateral terrestrial water fluxes are usually neglected. We estimated the contribution of vertical and lateral groundwater fluxes to the land surface water budget at a subcontinental scale, by modelling convergence of groundwater and surfacewater fluxes. We present a hydrological model of the entire Danube Basin at 5 km resolution, and use it to show the importance of groundwater for the surface climate.

The contribution of groundwater to evaporation is significant, and can be upwards of 30% in summer. We show that this contribution is local by presenting the groundwater travel times and the magnitude of groundwater convergence. Throughout the Danube Basin the lateral fluxes of groundwater are negligible when modelling at this scale and resolution. Also, it is shown that the contribution of groundwater to evaporation has important temporal characteristics. An experiment with the same model shows that a wet episode influences groundwaters contribution to summer evaporation for several years afterwards. This indicates that modelling groundwater flow has the potential to augment the multi-year memory of climate models.

1 Introduction

In the last decades, the importance of land-surface – atmosphere feedbacks in climate has been more and more recognized. Precipitation recycling, the process from local evaporation to local precipitation, is widely considered one of the important land-atmosphere interactions in the climate system (e.g., Trenberth, 1999; Brubaker et al., 1993; Koster et al., 2004; Bisselink and Dolman, 2009).

The strength of this feedback has been estimated in terms of rainfall recycling ratio (Trenberth, 1999) and coupling strength, where the latter can be estimated in terms of precipitation amounts (e.g., Koster et al., 2004; Dirmeyer, 2005) or rainfall probability (Lam et al., 2007). Key to precipitation recycling from an atmospheric perspective is evaporation. From a terrestrial perspective, runoff is the key process (Savenije, 1996), as all water that runs off the land surface, cannot contribute to evaporation and hence to precipitation recycling.

In this paper we take the compartmentalisation of water fluxes one step further. Just as the source of precipitation may be local (evaporation) or imported (by advection) (Trenberth, 1999), the source of evaporation may be local (from previous precipitation) or imported (by lateral transport). Terrestrial water has two major modes of lateral transport: surface water flow and groundwater flow. Both modes interact with soil moisture. Groundwater flows along a gradient that is usually dominated by the gradient in elevation, i.e. by topography. In flat terrains, in absence of topography-related gradients, groundwater is free to engage in lateral movements in any direction within aquifers, as gradients are dominated by gradients in aquifer downward (recharge) and upward (seepage, extraction, capillary rise) fluxes. So, it would be possible for groundwater to replenish episodic, local water shortages or to sustain a steady flux of water into regions that have a more persistent shortage of water. Surface water, on the other hand, flows along a predefined pattern (the river network), in a predefined direction (downstream). Sustained transport over large distances is normal, and contribution to the soil water in the land surface is possible via river – aquifer interactions.

Climate models suffer from a lack of "memory" in their land surface. Once a soil column has been completely dry or thoroughly wet, it carries no signal from past events. Persistences of over a year are seldomly seen. As groundwater flow is a slow process, it has been suggested that groundwater convergence may lead to persistence in surface climate (Bierkens and Van den Hurk, 2007; Maxwell and Miller, 2005; Maxwell and Kollet, 2008; Fan et al., 2007; Miguez-Macho et al., 2007; Anyah et al., 2008), although the effect of lateral flow has not yet been distinguished from the effect of in situ groundwater table dynamics (Yeh and Eltahir, 2005; Gulden et al., 2007; Ferguson and Maxwell, 2010; Fan and Miguez-Macho, 2010). On the one hand, a back-of-envelope calculation suggests that on the typical scale of a climate model, this effect should be small. On the other hand, the topology of the landscape could locally amplify the signal from groundwater convergence. The idea is that groundwater that has recharged in

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topographically higher areas flows towards (and converges to) lower exfiltration zones, mainly river valleys and wetlands, and sustains evaporation in the exfiltration zone either by capillary rise or by direct extraction of phreatophytes. Through the large inertia of groundwater systems wet periods from the past may then have an effect on increased evaporation in subsequent dry periods. By the evaporation–precipitation feedback (recycling) this may also affect warm season rainfall.

As lateral fluxes of groundwater (and surface water) are not represented in current climate models, the question arises how important this omission is for the reproduction of mass and energy balances of the land surface by climate models. As a first step to answering this question we employ a large-scale coupled groundwater–surfacewater model of the Danube basin to answer the following research questions.

1. What is the spatial and temporal contribution of groundwater to evaporation?

2. What fraction of groundwater-supported evaporation is local, and what fraction is imported (by river or groundwater convergence)?

3. What are the temporal (multi-year) connections in the groundwater contribution?

Studies that consider the first research question have been performed before (e.g., Bauer et al., 2006; Maxwell and Kollet, 2008; Miguez-Macho et al., 2007; Anyah et al., 2008). However, these studies were at meso-scale and considered regions where topography is the prime control of groundwater flow, while our study is at the regional climate scale, and focuses on large, relatively flat areas.

This paper is organized in five sections. The next section presents the Danube Basin as test bed for our research. Section 3 concerns the different compartments in our model and their integration, the climate forcing we used, and model calibration. In Sect. 4 we present the results and a discussion. In Sect. 5 we summarize the findings of our research and present general conclusions on spatial and temporal connections in the land surface water balance.

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2 The Danube Basin

The Danube Basin is an interesting test-bed for our analysis for several reasons. Recent studies suggested a strong soil-moisture precipitation feedback in parts of the basin (Seneviratne et al., 2006), regional climate models have shown a persistent dry bias in this region (e.g., Jacob et al., 2007; Kjellström et al., 2007) and the basin includes very large groundwater bodies. The Danube River Basin is the second largest river basin in Europe (after the Volga), covering around 800 000 km^2 in several countries and draining into the Black Sea (Regionale Zusammenarbeit der Donauländer, 1986).

The climate of the basin has a distinct W–E gradient. While the upper reach of the Danube, in the Western part of the basin, north of the Alps, has an atlantic influence, the middle and lower reach, in the eastern parts of the basin, have a more continental climate, with cold winters and dry summers. A reference hydroclimatology is given by Domokos and Sass (1990). The Pannonian Plain (region A in Fig. 1) is a region with very flat terrain. Quaternary lake and river deposits have a thickness up to 500 m, and both these and underlying deposits are large groundwater reservoirs. This plain is crossed by several rivers, of which the Danube (in its middle reach) and the Tisza are the largest. The Wallachian Plain (region B in Fig. 1) is a region with major groundwater reservoirs in the lower reach of the Danube. This plain has a history similar to the Pannonian Plain, with Quaternary uplift of the Karpathian mountains, regional basin subsidence and sedimentary aggradation. A major difference between the two regions is that the baselevel of the Wallachian Plain is determined by the Black Sea, its great oscillations contributing to formation of river terrasses and incised river valleys throughout this region (Radoane et al., 2003; Gilbrich et al., 2001).

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3 Modelling terrestrial water in the Danube Basin

3.1 Model framework and domains

The model is a modular, distributed, grid-based model, developed in the PCRaster environment (Wesseling et al., 1996; Karssenberg et al., 2007). We chose a 7 day timestep and a 5 km grid cell size. Figure 2 shows a schematic of the model set-up, in both the "steep" and the "flat" terrains. The "steep" terrains are all regions within the Danube Basin except the two recognized "flat" regions: the Pannonian and Wallachian Plains (A and B in Fig. 1).

3.2 Climate forcing

The model was forced by the 50 yr (1950–2000), 1° resolution, daily, global meteorological forcing dataset (Sheffield et al., 2006). The coarse resolution of the dataset introduces unwanted contrast at the edges of the climate grid cells (see Fig. 3, upper panel). Therefore, the data was spatially downscaled for our model domain by means of regression to altitude, as follows. If a climate variable Y correlates significantly with average altitude per grid cell $\bar{z}$ on the scale of the climate dataset (resolution ±100 km) in the land surface domain of Fig. 1, we assume that the regression coefficient β_1 in the relation $Y = \beta_0 + \beta_1\bar{z} + e$ is also valid on the model scale (resolution 5 km), so that we can use it as an environmental lapse rate. The estimated value of the climate variable at the model scale $\hat{y} = Y + \beta_1(z - \bar{z})$, where z is the altitude on the model scale. The resulting daily fields were temporally upscaled to the weekly time step of our model by taking the simple mean. The bottom panel of Fig. 3 shows a typical result of this downscaling. The contrast at the edges of the climate date grid cells is clearly reduced when compared to the original.

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3.3 Vegetation and snow cover

The snow pack intercepts all precipitation, whether in the form of snow or rain. Precipitation at temperatures 0 °C or below is assumed to be solid (snow), above 0 °C it is assumed to be rain. The snow pack can intercept rain only up to a certain fraction. If there is too much liquid water in the snow pack, it is removed as runoff. Liquid snow water is assumed to refreeze, allowing for more than one rainy episode per season.

A temperature index (degree-day) snowmelt method (see e.g., Hock, 2003) was used to model snow melt. Snow melt M_t is modeled as $M_t = f_m \cdot T^+ \Delta t$ where f_m is the snow melt factor, and T^+ is the cumulative positive difference between daily average temperature T_{avg} and melting threshold temperature T_0 during the time step Δt. The constant T_0 has a widely known "correct" value of 273 K but is calibrated (see Sect. 3.6) in concert with f_m to arrive at reasonable rates of snow cover disappearance in spring.

Snow evaporation or sublimation is a process that is notoriously difficult to model, as it depends on wind, radiation, snow albedo, snow compaction and several vegetation characteristics including snow interception characteristics (Pomeroy et al., 1998). As the necessary information is not available, snow evaporation is not included as a process in the model. Instead, we employ a simple correction factor at snow melt to avoid overestimating river discharge.

3.4 Soil water and evaporation

The soil water balance reads

$$\Delta S = P + M_t - E - R_{gw} - Q_r. \tag{1}$$

Precipitation P and snowmelt M_t add to the soil water budget; evaporation E, groundwater recharge R_{gw} and runoff Q_r subtract from the budget. The net effect of all these fluxes is ΔS, the change in soil water storage. All parts of the equation are fluxes of water (mass/area/time, simplified as length/time).

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The soil water flow q [m d^{-1}] is modelled after Campbell (1974), the conductivity k and the pressure head ψ [m] of a soil depends on its water content W [dimensionless].

$$q = -k\left(\frac{d\psi}{dz}+1\right) \tag{2}$$

$$\psi = \psi_{\text{sat}}\left(\frac{W}{W_{\text{sat}}}\right)^{-b} \tag{3}$$

$$k = k_{\text{sat}}\left(\frac{W}{W_{\text{sat}}}\right)^{2b+3} \tag{4}$$

We use the FAO Soil Map of the World (FAO, 1998) with the commonly used parameterset of Clapp and Hornberger (1978) to distinguish soil classes and arrive at estimates for W_{sat}, ψ_{sat}, k_{sat} and b (Braun and Schädler, 2005). Evaporation demand E_0 is computed using the standard FAO Penman Monteith method (Allen et al., 1998). The evaporation demand is met, in order of preference, by fluxes out of interception store E_{c}, out of soil water E_{swc}, and in flat areas by capillary rise out of the groundwater E_{gw} (Eq. 5). The potential flux out of interception store $E_{\text{c.pot}}$ during a timestep is limited only by the amount in store S_{c}. The potential flux out of soil water $E_{\text{s.pot}}$ is limited by the amount in store S_{s}, and by the conductivity of the soil. The potential flux out of the groundwater $E_{\text{gw.pot}}$ is limited by the amount of groundwater above a threshold level -5 m, which is determined by the relative groundwater level H_{rel} (groundwater level – surface level) and specific yield of the aquifer (taken equal to W_{sat}), and the conductivity of the unsaturated zone above the aquifer (taken equal to k), and spatially limited to the flat areas.

$$E = \begin{cases} E_0, & \text{for } E_o \leqslant E_{\text{c.pot}}, \\ E_{\text{c.pot}} + (E_0 - E_{\text{c.p}}), & \text{for } E_{\text{c.pot}} \leqslant E_0 \leqslant E_{\text{c.pot}} + E_{\text{s.pot}}, \\ E_{\text{c.pot}} + E_{\text{s.pot}} + \min(E_{\text{gw.pot}}, E_0 - (E_{\text{c.pot}} + E_{\text{s.pot}})), & \text{for } E_0 > E_{\text{c.pot}} + E_{\text{s.pot}}. \end{cases} \tag{5}$$

$$E_{\text{c.pot}} = S_{\text{c}}/\Delta t \tag{6}$$

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$$E_{\text{s.pot}} = \min(S_s/\Delta t, k) \tag{7}$$

$$E_{\text{gw.pot}} = \begin{cases} \min((-5 - H_{\text{rel}})W_{\text{sat}}, -k(\frac{\psi_{\text{sat}}-\psi}{-H_{\text{rel}}-1}+1)), & \text{for } H_{\text{rel}} \geqslant -5 \\ 0, & \text{for } H_{\text{rel}} < -5 \end{cases} \tag{8}$$

In flat areas, capillary rise that is not immediately consumed for evaporation is added to the soil, and is subsequently available for evaporation.

3.5 Groundwater and rivers

Groundwater in steep terrain has a contribution to the river discharge mainly as baseflow. We model the groundwater contribution to discharge by a linear reservoir. The baseflow at any timestep is given by $Q = S/\alpha$, with S the groundwater store and α the reservoir coefficient.

In the geologic setting of the Pannonian and Wallachian Plains, it is clear that the flat terrains contain the thick aquifers. In these areas, the groundwater flow is not primarily topography driven, and groundwater flow is two-dimensional. Therefore the groundwater is modelled by MODFLOW (Harbaugh et al., 2000). The deliniation of the two domains is based on topography, where the plains modelled by MODFLOW are two contiguous regions with less than 0.5% of slope, with constant-head boundary conditions. We use MODFLOW coupled into the model framework, as in Schmitz et al. (2009). We used spatially uniform aquifer properties, consistent with values obtained from Regionale Zusammenarbeit der Donauländer (1986). This also allows a double-check of the computed lateral fluxes by derivation of steady-state fluxes given the groundwater head at any timestep.

The surface water drainage network is obtained from a SRTM-derived digital elevation model (Jarvis et al., 2008). All rivers have an equilibrium width and incision depth according to Lacey's formula (see Savenije, 2003 for details and discussion). Rivers have interaction with aquifers (Sophocleous, 2002) so that they can locally recharge or drain the aquifer. The gradient between river stage and groundwater head is the

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driving force. We assume that a saturated connection between river and aquifer exists at all times.

3.6 Calibration

Using the topology of the basin, we calibrated the following key parameters sequentially. First, the degree-day factor and melting temperature of snow were calibrated, to arrive at reasonable rates of snow cover disappearance in spring. For the next steps, calibration was carried out using measured runoff data provided by GRDC. The model is not really exhaustively calibrated, but tuned until we arrived at a plausible and satisfactory set of parameter values. At each step, we computed runoff using a set of parameter combinations. Using the mean monthly discharge at the Danube Delta, we estimated the crop factors for the evaporation scheme, which also resulted in the need for a snow evaporation factor. Subsequently, using the mean monthly discharge at Bratislava, we estimated the reservoir constant of "steep groundwater". In the last step, using timeseries of discharge at the Iron Gate and at the Danube Delta, we estimated river bottom characteristics and aquifer properties. Figure 4 shows time series of calibrated versus measured discharge at the three measuring stations. In the upstream parts, shown by the Bratislava time series, the calibrated discharge reacts somewhat slower to changes in input than the measured discharge. In the middle and lower reaches, as shown by the Iron Gate and Danube Delta time series, respectively, modelled and measured discharges are in good agreement both by volume and by timing. We stress that we use this model only as a numerical laboratory to investigate the plausibility of groundwater contribution to the land surface water balance, and that discharge prediction was not a goal in constructing this model.

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4 Results and discussion

4.1 Groundwater contribution to evaporation

To assess the spatial and temporal contribution of groundwater to evaporation, it should be noted that this contribution has a direct and indirect component. In dry conditions, evaporation is possible directly from the groundwater. The indirect component is capillary rise: vertical flow from the groundwater table to the soil. The model keeps account of all fluxes in and out of the soil and thus the composition of soil water with respect to source is known. We assume perfect mixing of the soil water (or, equivalently, vegetation indifference with respect to water provenance) such that the relative contribution of groundwater to evaporation is equal to the relative content of groundwater-sourced soil water.

Figure 5 shows the importance of groundwater contribution to evaporation. In winter (lower left), groundwater does not contribute to evaporation, due to small evaporation demand, and due to snow cover. In all other seasons, groundwater contributes to evaporation significantly, in both the Pannonian and the Wallachian Plains. The patchy pattern of groundwaters contribution to evaporation is caused by the river network: where rivers are incised, the groundwater levels are more likely to stay below the interaction level of 5 m below the land surface.

4.2 Contribution of imported groundwater to evaporation

Our model simulates two modes of lateral transport that can possibly contribute to evaporation: transport by river and transport by groundwater flow. Groundwater is free to engage in lateral movements in any direction within the aquifers, so it would be possible to replenish localized shortages or sustain a steady flux of water into regions that have stronger coupling with atmosphere. To calculate the importance of these processes, we derive both groundwater velocity v [yr km^{-1} in a lateral direction] and convergence conv [$\mathrm{m\,d^{-1}}$ in the vertical] using the spatial distribution of groundwater level H and aquifer conductivity k and porosity n:

$$v = \frac{k\nabla H}{n} \tag{9}$$

$$\mathrm{conv} = \nabla^2 H \tag{10}$$

Figure 6 is a map of equilibrium groundwater travel velocity v, in $\mathrm{yr\,km^{-1}}$. This map shows that the time scale to transport water in the subsurface between adjacent cells (with a cell size of 5 km) is at minimum tens of years, and at maximum tens of thousands of years. We may note as an aside that atmospheric processes that transport water and share the same spatial scale and also interact with the land surface (e.g. cloud formation, storms, fronts) have typical time scales of hours to days, i.e. 4–6 orders of magnitude faster. The typical timescale is a telltale, but not sufficient to disprove the importance of lateral groundwater flow to the surface climate. For this, we also need to quantify the flux of water that the groundwater system makes available to the land surface.

Figure 7 shows that the typical magnitude of the groundwater convergence flux conv is in the order of $10 \times 10^{-7}\,\mathrm{m\,d^{-1}}$. At this rate, the groundwater convergence takes years to supply for one hour of summer evaporation. The expectation that lower-lying basins receive groundwater from the surrounding hills and mountains, is met by our simulation: along the boundaries of the two basins, export of groundwater is prevalent. The water pathways exist and the fluxes can be estimated, although they are of no importance to the surface climate in our model experiment.

It can be concluded that in the Danube region, the contribution of large-scale lateral groundwater flow to the land surface water balance, and therefore to the climate, is negligible. One caveat is that the modelled region in this study is relatively flat, so that the difference in groundwater storage between regions becomes the major gradient that drives the groundwater flow. Differences in groundwater storage in this region are determined by gradients in climate forcing, soil properties or vegetation properties at the appropriate scale. That is not to say that lateral groundwater flow plays no role in the land surface climate, but that at the scale of current climate models, it can be regarded as a local interaction.

The travelling times that we derived at a spatial resolution of 5 km indicate that the inclusion of lateral groundwater flow in deep aquifers under flat terrain only becomes useful when investigating at time scales of thousands of years. At those time scales the position of the land surface cannot be considered constant, as tectonics, sedimentary basin development and climate-related ice coverage change the landscape continually.

4.3 Temporal persistence in the coupled system

An additional model experiment was used to determine if multi-year memory exists in the coupled groundwater–soil model. The setup is as follows, and illustrated in Fig. 8. The model was run several times, each run starting with the same initial conditions. The reference run was forced by 10 years of unaltered climate forcing (Sheffield et al., 2006) (left black dot in Fig. 8) as a spin-up period. Then, we applied 10 years of cyclic median climate forcing, symbolized by the regular wave pattern, to arrive at a cyclic equilibrium state of the land surface. Then (right black dot in Fig. 8) unaltered climate forcing was applied for 1971, a very dry year in Europe. Subsequent runs each had a wet anomaly in the cyclic median climate forcing, symbolized by the grey ∼ on top of the regular wave, for each next run the anomaly was shifted one year back in time. The anomaly is constant throughout the year, so that the total precipitation during an anomalous year is at the 90th percentile of yearly precipitation in the dataset of Sheffield et al. (2006).

Figure 9 shows that several large parts of the Pannonian Plain and a few small areas in the west of the Wallachian Plain receive more groundwater for evaporation when the wet anomaly is one year before the dry summer. When the anomaly recedes in time, both the area and the magnitude of change in groundwater contribution to evaporation diminishes. The contrasting behaviour of the two regions is due to the fact that in most parts of the Wallachian Plain the equilibrium groundwater table is lower than 5 m below the surface, effectively prohibiting interaction between aquifer

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and land surface. The increased groundwater recharge during a single wet year is insufficient to raise the groundwater level above the interaction threshold of 5 m below the surface. The land surface and the aquifer stay uncoupled. In the Pannonian Plain, there are more regions where the increased recharge raises the water table above the interaction threshold, and also more regions were the equilibrium water table is above the interaction threshold even in absence of a wet anomaly. The wet anomaly is visible in the land surface water balance for up to 4 years.

5 Conclusions

Goal of this research was to investigate the importance of groundwater and groundwater convergence to the regional scale evaporation and through this on regional climate. To this end we built a coupled groundwater-soil moisture–surface water model of the Danube Basin, where land surface–precipitation feedbacks are expected to be significant.

Results show that groundwater contribution to dry season evaporation is significant in the two large groundwater basins considered, with relative contributions up to 30% and absolute area average rates over 1 mm d^{-1} in the Pannonian Plain. This analysis does not include the added effect on evaporation by irrigation from both groundwater and surface water, which may be significant.

Vertical groundwater flow (aquifers interacting with soil and rivers) is an important contributor to the land surface water balance, and should not be neglected in land surface models.

Travel time distributions are useful for including/excluding lateral fluxes in climate model experiments. At resolutions and time scales that are usual for climate models, lateral groundwater flow under flat terrain can be neglected. The travel times are too large and the fluxes too small to influence the land surface water balance and the surface climate. However, lateral subsurface flow takes place, and in a land surface

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model it can provide a realistic and efficient mechanism to close the model's water balance.

We also showed that large groundwater basins can be the cause of a significant multi-year local persistence to dry season evaporation, which is not included in current land surface models. The limited importance of lateral groundwater flow shows that these effects could easily be incorporated by replacing the leaky lower soil reservoir of a land surface model in flat sedimentary basins with a large capacity groundwater reservoir with zero bottom flux and the possibility of draining to the surface water only. However, compared to the costs of running an atmospheric model, running a groundwater model as part of the land-surface model is computationally cheap. So, apart from difficulties in parameterizing a large-scale groundwater model, there is no reason not to include groundwater dynamics in future land-surface components of regional or even global climate models. The added value would be a full closure of the coupled atmospheric-terrestrial water balance.

The groundwater component in this model significantly improves the persistence of the water cycle, regionally adding up to 5 years of delayed evaporation response to a wet episode.

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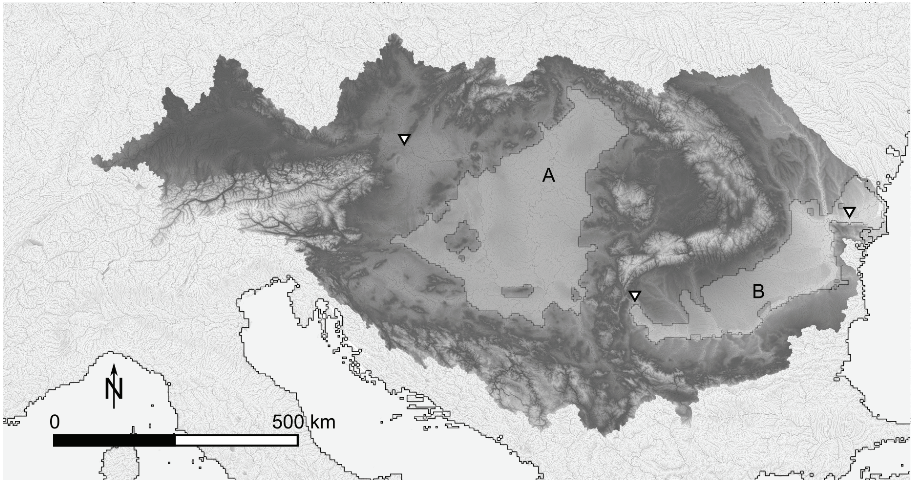

Fig. 1. Map of the Danube Basin. The regions labeled A and B are flat terrains where we modelled groundwater using MODFLOW. Region A is in the text referred to as the Pannonian Plain, region B the Wallachian Plain. The symbols ▽ mark the river discharge measurement stations Bratislava, Iron Gate and Ceatal Izmail (from West to East).

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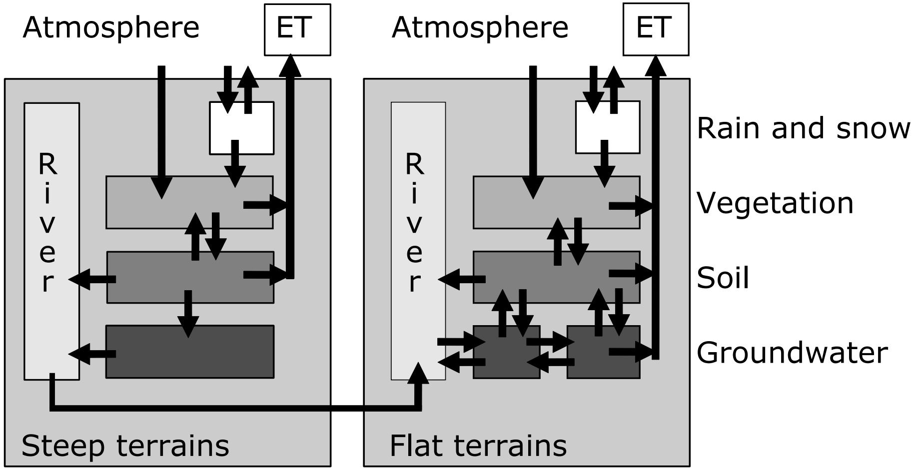

Fig. 2. Model schematics. In steep terrains (left) Groundwater contribution to runoff is modelled by a linear reservoir at each gridcell. In contrast, in flat terrains (right), groundwater level and flow is modelled with MODFLOW, allowing spatial interactions as well as (vertical) interactions with the land surface. The land surface components – vegetation, soil and surface water – are the same in both domains.

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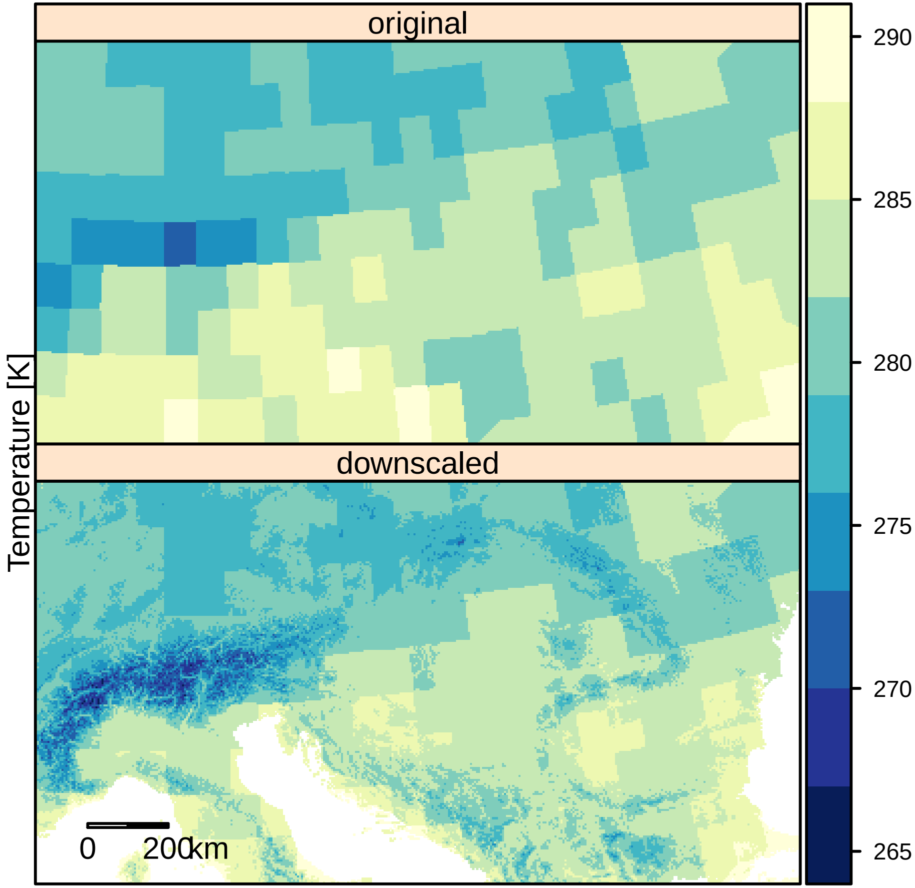

Fig. 3. Downscaling of climate variables. The panels have the same domain as Fig. 1. The upper panel shows the mean temperature on an arbitrary day, on the scale of the climate dataset (resolution ±100 km). The lower panel shows the downscaled temperature on the model scale (resolution 5 km).

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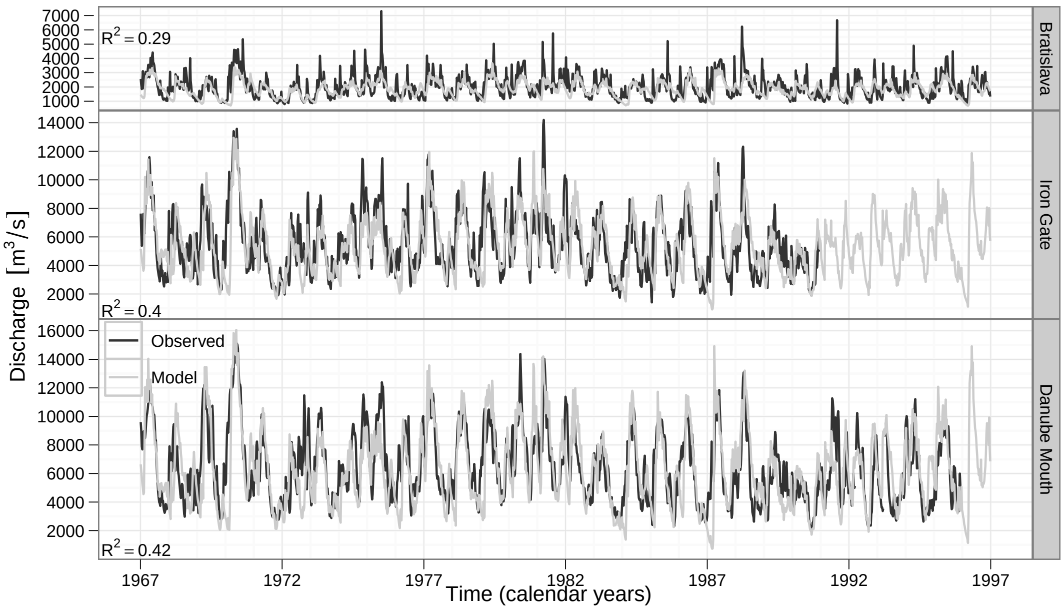

Fig. 4. Example hydrographs resulting from calibration. See Fig. 1 for the location of gauging stations Bratislava, Iron Gate and Danube Mouth (Ceatal Izmail).

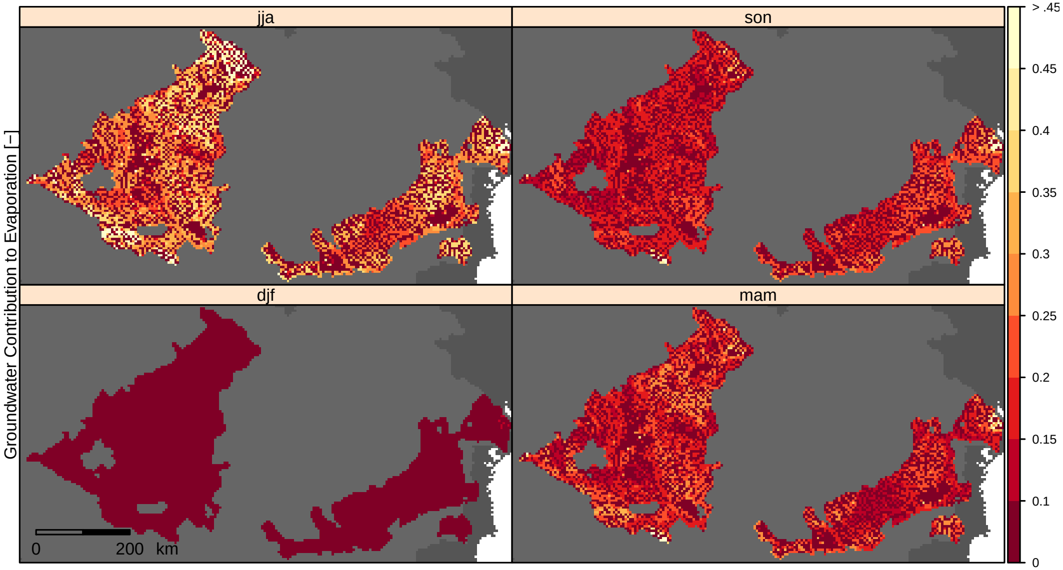

Fig. 5. Relative groundwater contribution to evaporation, per season. In winter (lower left), groundwater does not contribute to evaporation. From spring to late fall, there is a significant contribution of groundwater to sustain evaporative fluxes.

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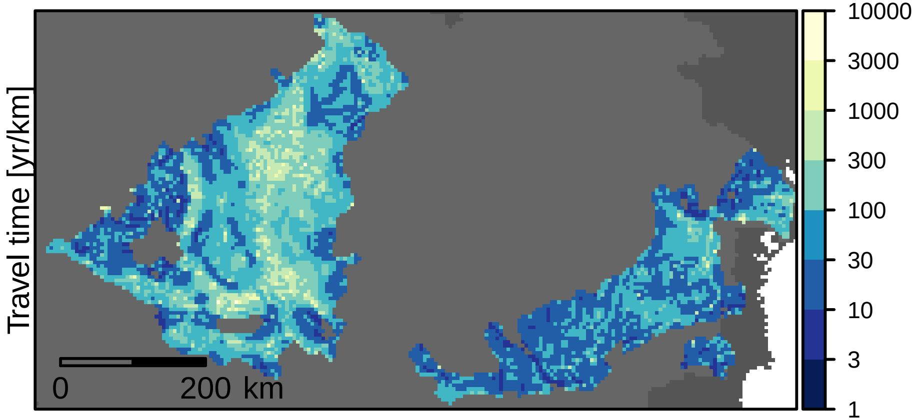

Fig. 6. Travel time of groundwater, in yr km^{-1}.

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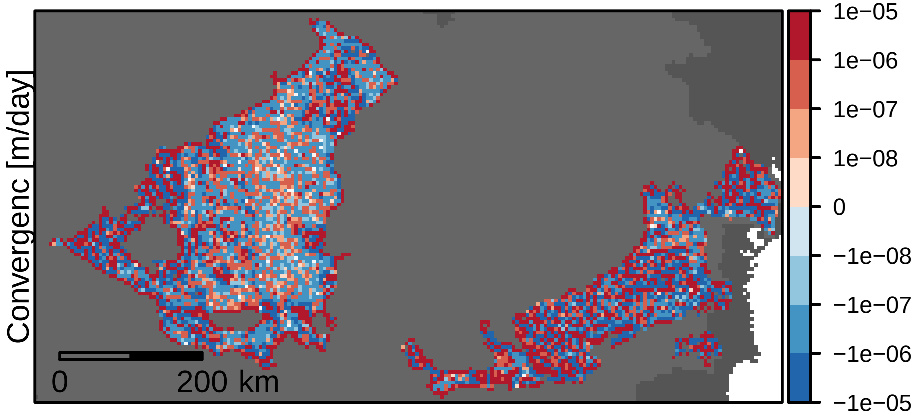

Fig. 7. Magnitude of groundwater convergence, in m d^{-1}. Positive (red) values are divergent areas, where there is a net groundwater export. Negative (blue) values are convergent areas, with a net groundwater import.

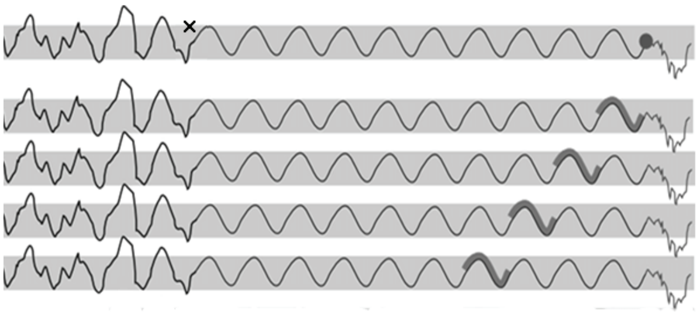

Fig. 8. Climate forcing sequences for estimating persistence in the coupled system. Irregular pattern denotes unaltered climate forcing (left, and right). At the "x" starts the cyclic median climate forcing. The grey ~ on top of the regular wave symbolizes a wet anomaly. The grey dot is symbol for the year 1971.

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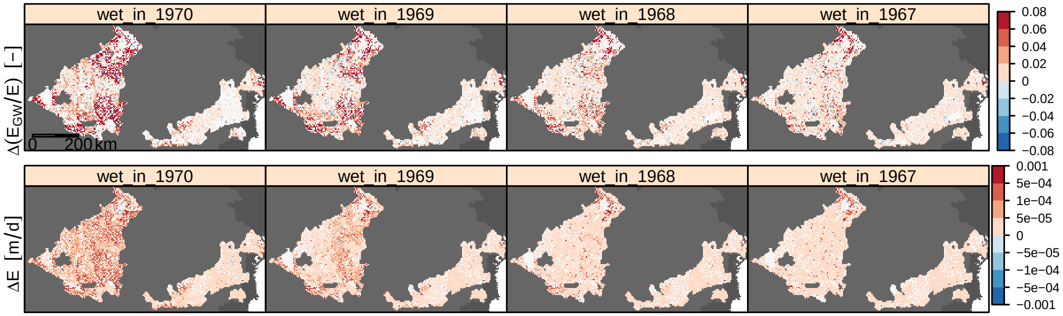

Fig. 9. Temporal persistence of summer evaporation and groundwater contribution to summer evaporation. Shown here is the difference in evaporation ΔE and the difference in groundwater contribution to evaporation $\Delta(E_{gw}/E)$ in a dry summer (1971) after 10 years of cyclic median climate forcing, compared to the same forcing except supplemented with a wet anomaly in the year just before (wet in 1970), two years before (wet in 1969), etc.