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Key Points:

- Hillslope flow sustains higher soil moisture, enhancing evapotranspiration and precipitation rates and cooling down the air temperature
- Hillslope flow slightly attenuates global warming
- At regional scale, hillslope flow attenuates climate change trends of all hydrological variables except for evapotranspiration increases

Supporting Information:

Supporting Information may be found in the online version of this article.

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Influence of Hillslope Flow on Hydroclimatic Evolution Under Climate Change

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Abstract We analyzed the influence of hillslope flow on projections of climate change by comparing two transient climate simulations with the IPSL climate model between 1980 and 2100. Hillslope flow induces a reorganization and increment of soil moisture (+10%), which increases evapotranspiration (+4%) and precipitation (+1%) and decreases total runoff (−3%) and air temperature (−0.1 °C) on an annual average over land for 1980–2010 when compared to simulation not representing hillslope flow. These changes in land/atmosphere fluxes are not homogenous and depend on regional climate and surface conditions. Hillslope flow also influences climate change projections. On average over land, it amplifies the positive trend of soil moisture (+23%), evapotranspiration (+50%), and precipitation (+7%) and slightly attenuates global warming (−1%), especially for daily maximum air temperature. The role of hillslope flow in supporting surface/atmosphere fluxes is more evident at a regional scale. Where precipitation is projected to decrease, hillslope flow is shown to attenuate the related declines in evapotranspiration, precipitation, and total runoff, regardless of aridity conditions and mean air temperature. Where precipitation is projected to increase, hillslope flow amplifies evapotranspiration enhancement but attenuates the increase in precipitation and total runoff. Warming is generally attenuated, especially in semiarid and cold areas, and humid and warm/temperate regions, but the signal is weak. These results demonstrate the role of hillslope flow in enhancing water and energy fluxes between the surface and the atmosphere. They also suggest that including hillslope flow in climate models would weaken the projected intensification of hydrological extreme events.

Plain Language Summary We analyze how the flows caused by topography, called hillslope flow, affect the evolution of climate using simulations from a climate model. Results show that hillslope flow increases soil moisture in the valleys. More soil moisture enhances evapotranspiration and precipitation and decreases total runoff and air temperature for the period 1980–2010. But the increase in water exchanges between land and atmosphere is not homogenous. In the future, hillslope flow amplifies positive trends of climate change for soil moisture, evapotranspiration and precipitation, while global warming is minimally slowed. The role of hillslope flow in sustaining exchanges of water and energy in the future is most evident in the regions. Where precipitation decreases in the future, evapotranspiration and precipitation declines are less intense when hillslope flow is included, and that is the case for the decline in total runoff as well. In regions where precipitation increases in the future, evapotranspiration increases faster, but precipitation and runoff increases slower. Effect of hillslope flow on warming is weak, but in general, the air warms up more slowly, especially in both semiarid/cold regions, and humid and warm/temperate regions. The results highlight the role of hillslope flows in increasing water exchange between the surface and atmosphere

1. Introduction

There is strong evidence that soil moisture plays a critical role in the evolution of climate (Seneviratne et al., 2010). For instance, at the seasonal scale, the GLACE project (Koster et al., 2006) revealed that soil moisture anomalies in the Northern Hemisphere, the Sahel and equatorial Africa can heavily affect seasonal values of precipitation and temperature (Koster, 2004). In Europe, soil moisture plays a role in the northward propagation of Mediterranean drought and in the occurrence of extreme summertime temperatures (Quesada et al., 2012; Zampieri et al., 2009); preexisting dry soil conditions are a necessary prerequisite for the occurrence of heat waves (Horton et al., 2016; Perkins, 2015). In the United States, irrigation-induced changes in soil moisture could help explain part of the modeling bias in summer temperature and precipitation (Al-Yaari et al., 2019; Barlage et al., 2021).

Further efforts in the GLACE-CMIP5 project analyzed the feedbacks of climate change and land-atmosphere coupling. Early results revealed that the projected decrease in soil moisture would result in a stronger decrease in evapotranspiration and precipitation and a higher increase in temperature at the end of the 21st century (Seneviratne et al., 2013). Other results pointed to feedback between decreasing soil moisture and land surface temperature, relative humidity, and precipitation, thus explaining an increase in aridity under climate change (Berg et al., 2016). By the end of the 21st century, in some regions, decreasing trends in soil moisture due to climate change can also lead to further increases in the intensity, frequency, and duration of temperature extremes and in dry precipitation extremes (Lorenz et al., 2016).

For land-atmosphere coupling, knowledge gaps exist regarding the effect of topography heterogeneity on energy and water fluxes between land and atmosphere. The topographic gradient induces a hillslope flow of surface water and groundwater between the upland area and the lowland valley (Fan et al., 2019), which can create a wetter riparian wetland (Fan & Miguez-Macho, 2011). The role of groundwater is especially important for hillslope flow, because groundwater is the largest continental reservoir (Gleeson et al., 2016), and its slow flow influences a longer soil moisture memory in areas with shallow water tables (Cuthbert et al., 2019; Gleeson et al., 2011; Martinez et al., 2016a, 2016b; Martínez-De La Torre & Miguez-Macho, 2019). In water-limited regions, wetter soil induced by groundwater increases evapotranspiration (ET) (Fan et al., 2019; Maxwell & Condon, 2016). Higher ET directly affects surface water and energy budgets and can lead to increased precipitation as response from the atmosphere (Lo & Famiglietti, 2011; Wang et al., 2018).

Modeling is a good option for disentangling the link between hillslope flow and land-atmosphere fluxes. Coupling a land surface model (LSM) to a general circulation model (GCM) allows us to explore the evolution of the land and atmospheric components at the global (Wang et al., 2018) and regional scales (Anyah et al., 2008; Campoy et al., 2013; Fan et al., 2007). In LSMs, the representation of groundwater storage and its contribution to river discharge is now commonly included, with multiple approaches from a simple reservoir to physically based representations with water table dynamics (Gleeson et al., 2021). If implemented at a high enough spatial resolution, the latter approaches allow one to explicitly simulate hillslope flow and its influence on soil moisture heterogeneities, but they remain very rare in climate models, and only in regional ones (Anyah et al., 2008; Furusho-Percot et al., 2019). In coarse resolution LSMs classically coupled to atmospheric models, hillslope processes remain a challenge (Clark et al., 2015). Most attempts make use of the TOPMODEL formalism (Band et al., 1993; Beven & Kirby, 1979), either in a diagnostic mode (Gedney & Cox, 2003), or with a full coupling of subgrid water table depth distribution with the ones of soil moisture, runoff, and ET (Ducharne et al., 2000; Koster et al., 2000; Walko et al., 2000). These models, however, may overestimate the moistening of soils by capillary fluxes from the water table, as it is assumed to always be shallow in TOPMODEL (Beven et al., 2021; Gascoin et al., 2009). Interactions with deeper groundwater systems can also be described in climate models, usually via 1D capillary fluxes to the soil, which overlooks the modulations by small scale (subgrid) topography. The most comprehensive description of groundwater-soil moisture interactions in a global climate model is presently offered by the ISBA-CTRIP LSM (Decharme et al., 2019)), combining 2D horizontal groundwater flow between grid cells with vertical capillary rise in the fraction of each grid cell with the lowest elevation.

Therefore, the feedback of hillslope flow on climate change projections is not clear. Climate change, especially warming, is expected to have an impact on groundwater recharge and storage (Markovich et al., 2016; Smerdon, 2017; Wu et al., 2020). Such changes in groundwater storage and hillslope flow may support higher ET rates in transition zones between wet and dry climates for some time, but not indefinitely (Condon et al., 2020). However, the evidence provided is regional, and the modeling efforts overlook dynamical interactions with the atmosphere. Because land-surface coupling has an impact on the projection of climate change, the reorganization of soil moisture induced by hillslope flow may play a role in projections of some hydroclimatic variables on a global scale.

Here, we present evidence of the influence of hillslope flow on global climate change projections of hydroclimatic variables. We contrast the results of two transient land-atmosphere simulations for the period 1979–2100, one of which uses the default land surface representation of the ORCHIDEE LSM (Section 2.1.1), while the other includes a novel subgrid hillslope flow parameterization (Section 2.1.2). As described in Section 2.2, we use LMDZOR, the coupled land-atmosphere component of the IPSL-CM6 climate model from the Institut Pierre

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Simon Laplace (Boucher et al., 2020; Cheruy et al., 2020). First, we focus the analysis on the historical period (1980–2010) to explore the effects of hillslope flow on yearly and seasonal average values in terms of both sensitivity and realism against observations (Section 3.1). We then explore the effect of hillslope flow on climate change trends, which we call hillslope flow modulation. To calculate the hillslope flow modulation, we first estimate the long-term trend for each simulation during the 21st century; then the trend of the difference of the two simulations at the global scale in Section 3.2; and finally, the spatial distribution of the hillslope flow modulation in Section 3.3. To further analyze hillslope flow modulation at the regional scale, we use a simple climate classification inspired by the Köppen-Geiger and Thornthwaite classifications (Beck et al., 2018; Feddema, 2005) in Section 3.4. The latter allows for defining which regions are prone to hillslope flow modulation and what is the type of modulation in these areas. In Section 4, we discuss the main limitations of our results, and we close in Section 5 with the main conclusions and perspectives.

2. Materials and Methods

2.1. Description of the Land Surface Model

2.1.1. ORCHIDEE-REF

ORCHIDEE (ORganizing Carbon and Hydrology in Dynamic EcosystEms) is a process-based model that describes the fluxes of mass, momentum, and heat between the surface and the atmosphere (Krinner et al., 2005). The version used here as a reference (called ORCHIDEE-REF in the following) corresponds to version 2.0, included in the IPSL-CM6 climate model for CMIP6 simulations and described in Cheruy et al. (2020), Boucher et al. (2020), and Tafasca et al. (2020). Here, we summarize the main characteristics of the model. In Section 2.1.2, we summarize the hillslope flow parametrization.

In each grid cell, vegetation is represented by a mosaic of up to 15 plant functional types (PFTs), including bare soil. Each PFT is characterized by a set of parameters (Boucher et al., 2020; Mizuochi et al., 2021), and fractions are described by the LUHv2 data set (Lurton et al., 2020). Plant phenology is controlled by the STOMATE module, which couples photosynthesis and the carbon cycle and computes the evolution of the leaf area index (LAI) (Krinner et al., 2005). It means that CO_2 influences plant growth and phenology. PFTs are grouped into three soil columns according to their physiological behavior: high vegetation (forest, eight PFTs), low vegetation (grasses and crops, six PFTs), and bare soil. A separate water budget is calculated independently for each soil column, in order to prevent forest PFTs from depriving the other PFTs of soil moisture but within each soil column, the uptake of water for transpiration considers the root distribution of the corresponding PFT (de Rosnay et al., 2002). In contrast, the energy balance is calculated for the whole grid cell (Boucher et al., 2020).

Evapotranspiration is represented by a classical bulk aerodynamic approach with four subfluxes: snow sublimation, interception loss, bare soil evaporation, and transpiration. The first two proceed at potential rates and originate from the snow-covered area and from the PFT fraction effectively covered by foliage. Bare soil evaporation is limited by upward water diffusion. Transpiration originates from the PFT fractions effectively covered by foliage, with no intercepted water, and is controlled by stomatal resistance, which depends on soil moisture and vegetation parameters.

Vertical soil water flow is represented by a 1-D Richards equation (Figure 1a) coupled to a mass balance (de Rosnay et al., 2002), and soil depth is set to 2 m, here discretized into 22 layers to finely model lower layers implicated in drainage (Campoy et al., 2013; Wang et al., 2018). Surface infiltration is limited by the hydraulic conductivity of surface layers and is represented by a sharp wetting front based on the Green and Ampt model (D'Orgeval et al., 2008; Tafasca et al., 2020). The resulting increase in top soil moisture is delivered to the Richards redistribution scheme as a boundary condition, while the lower one consists in free drainage equal to the hydraulic conductivity of the deepest node. Lateral fluxes between cells are neglected (Campoy et al., 2013; de Rosnay et al., 2002).

Soil is assumed to be homogeneous inside every grid cell and represented by the dominant USDA soil texture, as taken from the Zobler (1986) map. Soil parameters are a function of the soil texture, following Carsel and Parrish (1988), while the unsaturated values of hydraulic conductivity and diffusivity depend on soil moisture using the Van Genuchten-Mualem model (Mualem, 1976; van Genuchten, 1980). Even if the soil texture is assumed to be uniform inside every grid cell, the saturated hydraulic conductivity decreases with depth, following D'Orgeval et al. (2008).

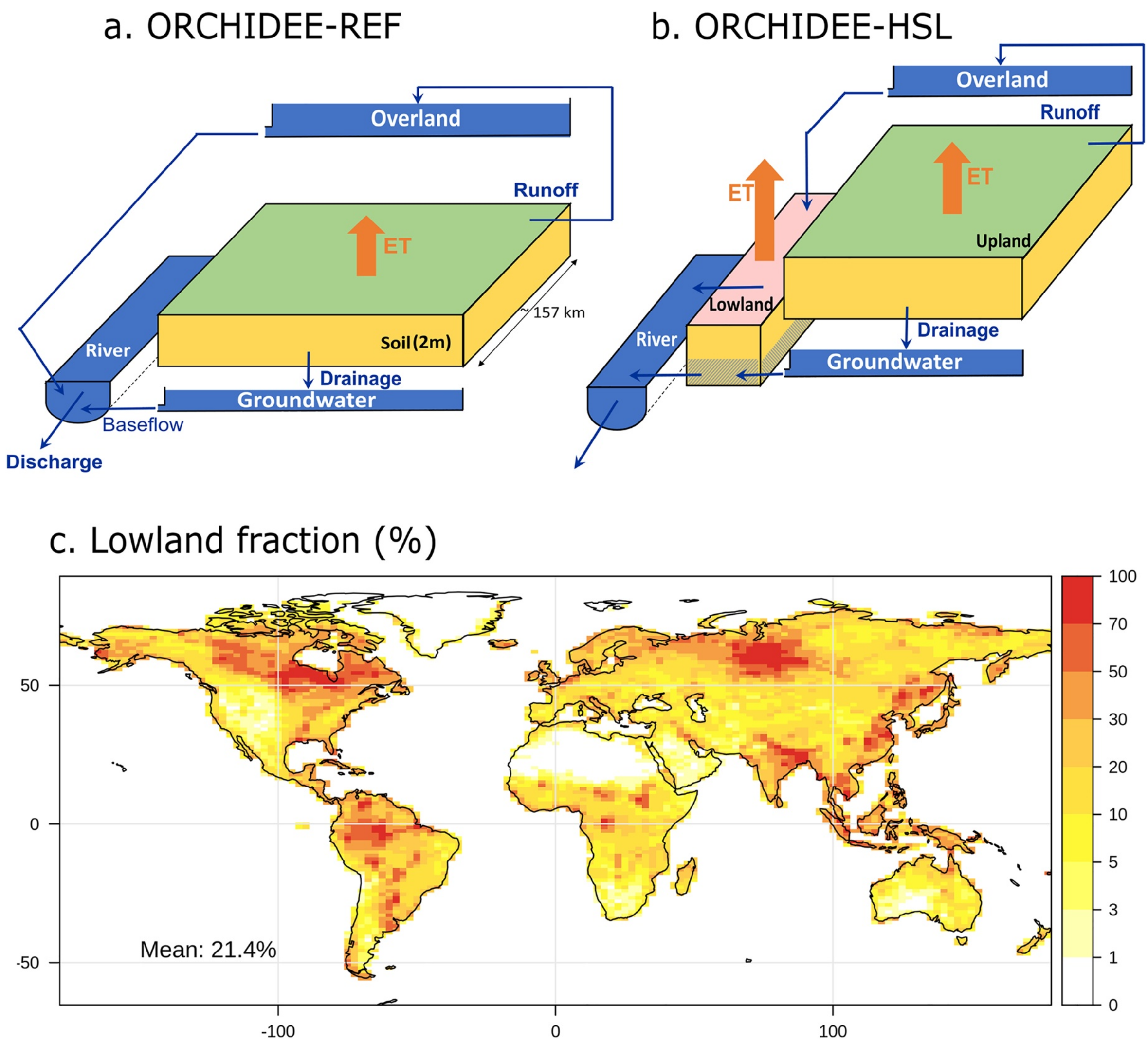

Figure 1. Schematization of the main hydrological fluxes in ORCHIDEE-REF (a) and ORCHIDEE-HSL (b), blue arrows represent surface and subsurface flows, orange arrows represent fluxes to the atmosphere. Map of lowland fraction from Tootchi et al. (2019) interpolated to ORganizing Carbon and Hydrology in Dynamic EcosystEms (ORCHIDEE) resolution (c), corresponding to 2.5 × 1.3° grid size.

The routing scheme (Figure 1a) transfers surface runoff and drainage from land to the ocean through a cascade of linear reservoirs (Guimberteau et al., 2012; Ngo-Duc et al., 2007). Each grid cell is split into subbasins according to a flow direction map based on the work of Vörösmarty et al. (2000), enhanced over the polar regions by Oki et al. (1999) with a resolution of 0.5°. Every subbasin includes three reservoirs (Figure 1a), corresponding to groundwater, overland storage, and stream storage, with decreasing residence times (Ngo-Duc et al., 2007). The groundwater reservoir collects drainage from the soil column, while the overland reservoir collects surface runoff. There is no feedback from these two reservoirs on soil moisture. In particular, there is no capillary rise from the groundwater reservoir, so that groundwater cannot influence the atmosphere through the soil column. These two reservoirs are internal to each subbasin, and they both feed the stream reservoir, which also collects streamflow from the upstream subbasins, hence contributing to large-scale routing across subbasins and grid cells.

Eventually, the surface energy and water budget are computed at the same time step as the atmospheric model, that is, 15 min. We impose the same time step on the routing scheme for consistency with ORCHIDEE-HSL. In contrast, the carbon and plant phenology processes in STOMATE are solved with a daily time step.

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2.1.2. ORCHIDEE-HSL

ORCHIDEE-HSL describes the effect of hillslope flow along topography at the subgrid scale. To this end, it introduces a "lowland" fraction, which buffers the flow between the upland fraction and the river system (Figure 1b) and has its own water budget, while the remaining "upland" fraction behaves like the soil of the entire grid cell in the reference version. Surface runoff and drainage replenish overland and groundwater reservoirs, which now feed the lowland fraction instead of the stream reservoir, but if there is no lowland fraction within the grid cell, overland and groundwater flows pass directly to the river, as in ORCHIDEE-REF. The lowland fraction can thus be seen as a topographically driven riparian wetland (Fan & Miguez-Macho, 2011), fed by the convergence of both surface water and groundwater, with a potentially higher evapotranspiration than the upland fraction. It is worth noting that this approach is equivalent to the use of the representative hillslope concept (Fan et al., 2019; Swenson et al., 2019), and is suitable for large-scale modeling efforts with redistribution at subgrid scale, but not for high-resolution simulations, for which intercell flow (2D or even 3D) is necessary (Felfelani et al., 2021; Markovich et al., 2016).

In contrast to many LSMs describing topographically driven wetlands with a variable area, such as in TOPMODEL (Band et al., 1993; Beven & Kirby, 1979), we suppose here for simplicity that the lowland fraction remains constant in a grid cell, and only its soil moisture changes with time. In this framework, the lowland fraction is prescribed from a 500-m resolution global-scale wetland map recently designed for this purpose (Tootchi et al., 2019). It combines open-water and inundation imagery and high-resolution groundwater modeling (Fan et al., 2013) and is interpolated to the ORCHIDEE resolution according to Section 2.2 (Figure 1c).

Even if some vegetation types may be favored/prevented in lowland areas, this is overlooked due to the lack of guiding rules to do otherwise (Fan et al., 2019). Therefore, the lowland fraction has the same PFT composition as the upland fraction (and as the full grid cell in the reference configuration). A given PFT, however, undergoes weaker water stress in the wetter lowland fraction, which therefore produces higher transpiration and a higher LAI. Bare soil evaporation is also enhanced by the higher soil moisture in the lowland fraction.

In this fraction, the overland flow from the upland fraction is added to throughfall and snowmelt. Depending on soil moisture and hydraulic conductivity, the resulting amount of water can either infiltrate into the soil or produce surface runoff, which directly flows to the river. The flow from the groundwater reservoir is injected at the bottom of the soil column, which is supposed to be impermeable to vertical drainage, so that the deep layers can gradually saturate, thus forming a water table that drains horizontally to the stream reservoir as baseflow. The water table is practically defined as the top of the uppermost saturated soil layer when starting from the impermeable soil bottom. If the whole soil column becomes saturated, the excess water will add to surface runoff and flow into the river (Tootchi, 2019). The soil depth is kept at 2 m in the lowland fraction, and it is discretized into 22 layers (with increasing height from 1 mm in the top layer to 12.5 mm in the eighth layer and all layers below) to accurately simulate the water table depth and the overlying soil moisture gradients and resulting water fluxes according to the Richards equation (Campoy et al., 2013). This discretization is also imposed for consistency in the upland fraction and in the full grid cell of ORCHIDEE-REF.

Baseflow Q_{base} (mm/s) to the streams comes from the lowland fraction. Following Darcy's law, it originates from the saturated layers below the water table, and is given by a solution of the long-term linearized Boussinesq equation at the catchment scale (Brutsaert, 2005; Tootchi, 2019) in equivalent water depth:

$$Q_{base} = \sum_{i=wtl}^{22} q_i = E_F * \frac{\pi^2}{4} * \frac{\Delta h}{B} * \frac{1}{B} * \sum_{i=wtl}^{22} \Delta z_i * K_i \quad (1)$$

where E_F [–] is the exchange factor multiplied by $\frac{\pi^2}{4}$, which accounts for variations of hydraulic conductivity along horizontal direction, anisotropy and riverbed clogging, based on water table shape (Brutsaert, 2005). The term $\frac{\Delta h}{B}$ corresponds to the mean gradient along hillslopes in the lowland fraction. It depends on Δh [L], the height of water table above the bottom of the soil column, that is, 2 m depth, and on B [L], the mean aquifer breadth from the streams to the divides. This term is equal to $\frac{1}{2*\delta}$, where δ [L^{-1}] is the drainage density in the lowland fraction, assumed here to be the same as in the entire grid cell by lack of specific information. The term $\Delta z_i * K_i$ corresponds to the layer transmissivity, dependent on the layer thickness Δz_i [L] and the layer hydraulic conductivity K_i [L/T]. Finally *wtl* is the layer that contains the water table.