Hydrology and Earth System Sciences Discussions

What do moisture recycling estimates tell? Lessons from an extreme global land-cover change model experiment

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Abstract

Moisture evaporated from the continents (recycled moisture) contributes up to 80% to the total atmospheric moisture content and, hence, precipitation in some regions. Recycling estimates are traditionally used to indicate a region's rainfall-dependence on land-surface evaporation. Accordingly, recycling estimates are employed to deduce the hydrological consequences of land-cover change. However, moisture is not a passive but an active constituent of the atmosphere. Recent studies indicate that at small scales (up to 1000 km) local to regional evaporation-precipitation coupling by far dominates the atmospheric precipitation response, while the water-balance effect from moisture recycling in the traditional sense seems to be of minor importance. The value of moisture recycling estimates as indicator for consequences of land-cover change is therefore questionable. However, since atmospheric moisture is still subject to mass conservation, the relevance of moisture recycling may come into play at the continental scale.

To explore the relevance of recycling estimates regarding land-cover change at the continental scale, we conduct two global experiments with an atmospheric general circulation model: (I) with present-day conditions and (II) with extreme land-cover change conditions, namely with totally suppressed continental evaporation. Using the simulated fields of moisture, wind, and evaporation from the present-day experiment, we quantify continental moisture recycling with a vertically integrating tracing scheme. We then compare the computed recycling patterns with the hydrological changes that follow the suppression of continental evaporation.

While under present-day conditions the fraction of recycled moisture increases from continental upstream to downstream regions with respect to the prevailing winds, the suppression of continental evaporation leads to severe precipitation loss in almost all continental regions, no matter if situated upstream or downstream. Over the ocean the hydrological response is ambigious, even where under present-day conditions large fractions of the atmospheric moisture stem from continental evaporation. This suggests

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that continental moisture recycling can not act across large ocean basins. Over land the absence of evaporative cooling at the surface leads to substantial warming which acts to suppress precipitation. In large parts of the continents the precipitation decrease compensates for much of the missing evaporation, such that the continental
5 moisture-sink is not much amplified. Consequently, the atmospheric moisture content is not systematically reduced in the evaporation-free experiment, as would be necessary for the traditional moisture recycling mechanism to be active. Noteworthy exceptions are continental regions that are substantial moisture sources for some time of the year, first of all tropical wet-dry climates during the dry season. Apart from these
10 exceptions, our results challenge the relevance of moisture recycling estimates for the hydrological consequences of land-cover change even at the continental scale.

1 Introduction

Source-target relations of completely passive atmospheric trace gases provide straight-forward information on consequences for the target region one expects from modified
15 emissions at the source region: changes at the source region result in predictable changes at the target region. But in how far does the situation change if the considered gas is not passive but an important active constituent of the Earth's atmosphere? What do source-target relations tell when it comes to water?

One of the most fundamental properties differentiating land from ocean is that land
20 can be dry. In contrast to ocean evaporation, land-surface evaporation is not supplied by an essentially inexhaustible water reservoir. Land-surface evaporation is limited by the amount of water supplied by precipitation and, hence, stays usually below potential evaporation. Also, even in most of those regions where annual-mean potential evaporation exceeds annual-mean precipitation, some of the water supplied by precipitation
25 runs off to the ocean basins via rivers due to gravity. This water loss is particularly pronounced where the atmosphere's supply of moisture (precipitation) and demand for moisture (potential evaporation) are markedly displaced in time. Where runoff occurs

the land-surface is a sink for atmospheric moisture, suggesting that air flowing over land successively dries out. Land surface characteristics including vegetation type and water management may influence the degree of moisture depletion of the passing air by controlling the fraction of precipitation that reevaporates instead of running off. The moisture supply of some continental region should therefore depend on the hydrological properties of the land-surface in the upstream regions. Hence, land-cover change activities in upstream regions should influence downstream precipitation. The depicted mechanism is commonly referred to as continental moisture recycling or precipitation recycling. In this paper we refer to this moisture-budget related mechanism as the traditional approach.

In traditional moisture recycling studies, the extent to which precipitation in a region depends on moisture recycling has been linked directly to the fraction of recycled moisture in total moisture (recycled moisture fraction, RMF, also referred to as recycling ratio), where total moisture can be vertically integrated atmospheric moisture (VIM) or precipitation. Therefore, these studies commonly aimed at estimating RMFs. Early studies on this issue aimed at estimating the contribution of evaporation from a particular region to precipitation inside the same region (e.g. Benton et al., 1950; Budyko, 1974; Lettau et al., 1979; Brubaker et al., 1993; Eltahir and Bras, 1994; Savenije, 1995a; Trenberth, 1999; Burde and Zangvil, 2001; Fitzmaurice, 2007). These studies are based on bulk recycling models that relate horizontal moisture influx and land-surface evaporation. For example, the bulk recycling equation applied by Brubaker et al. (1993) reads

$$\text{RMF}_\text{reg} = \frac{E_\text{reg}}{E_\text{reg} + 2 \cdot F^\text{in}_\text{reg}} \tag{1}$$

where RMF_reg is the region's mean fraction of recycled moisture in precipitation, F^in_reg is the region's horizontal moisture influx, and E_reg is the region's land-surface evaporation. The individual bulk recycling models differ slightly from each other, with more recent models typically designed to relax the underlying assumptions. They are easily

applicable to available data on moisture fluxes and land-surface evaporation. However, they have to employ simplifying assumptions that are hard to justify (see Fitzmaurice, 2007 for details). Also, the results depend strongly on the size and the shape of the considered region, because E_{reg} scales with the region's area and F_{reg}^{in} scales with its diameter perpendicular to the prevailing wind direction. Because of this scale dependence the RMF_{reg} measure is of limited use for the intercomparison of moisture recycling estimates between different regions.

Continental RMFs do not suffer these scale issues simply because the definition of the considered region is free from arbitrariness: the global land-surface. However, although bulk recycling models can in principle be applied to the global land-surface, the assumptions made in the model derivations are not appropriate given the size, shape, and heterogeneity of the continents. Also, with growing complexity of the considered region, one is interested in spatially resolved RMFs rather than one single mean value. Alternatively, continental RMFs can be computed from wind, moisture, and evaporation data by diagnostic tracing of water that is tagged according to its origin. This approach has been adopted by Numaguti (1999), Bosilovich et al. (2002), Yoshimura et al. (2004), and van der Ent et al. (2010). While Numaguti (1999) and Bosilovich et al. (2002) used pure general circulation model (GCM) data to trace moisture 3-dimensionally, Yoshimura et al. (2004) and van der Ent et al. (2010) used 2-dimensional horizontal moisture fluxes derived from reanalysis data.

Yoshimura et al. (2004) and van der Ent et al. (2010) applied the well-mixed assumption, reducing the problem to two spatial dimensions. This assumption implies that the atmosphere is vertically well-mixed with respect to RMFs, which is one of the simplifications also employed in most of the bulk recycling models (see Fitzmaurice, 2007). The approach taken by Numaguti (1999) and Bosilovich et al. (2002) does not require the well-mixed assumption, because the vertical moisture exchange is resolved explicitly. Despite this difference the authors of the four studies cited above found similar RMF estimates. As expected, RMFs increase from upstream continental regions to downstream continental regions, for example from west to east over North America and

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Eurasia, and from northeast to southwest over Amazonia. RMF maxima are around 60% in the tropics (year-round), and even 80% in East Central Asia (during northern summer).

The question remains what the significance of moisture recycling estimates actually is. Based on mere moisture-budget considerations, as employed in the traditional moisture recycling studies, it is reasonable to assume a close relation between RMFs and a region's sensitivity to upstream evaporation. This expectation is actually the reason for the interest in RMF estimates, because such a sensitivity measure could help to predict the hydrological consequences of certain land-cover change activities. However, more recent studies show that the hydrological state of the land-surface does not only influence the atmosphere's moisture budget in a spatio-temporally integrating manner. The hydrological state of the land-surface exerts strong influence on local to regional atmospheric conditions and, hence, precipitation (e.g. Shukla and Mintz, 1982; Rowntree and Bolton, 1983; Betts et al., 1996; Findell and Eltahir, 1997; Schaer et al., 1999; Pal and Eltahir, 2001; Koster et al., 2002, 2004, 2006). In a regional modelling study over Europe Schaer et al. (1999) conclude from moisture-budget calculations that "the simulated sensitivity (to soil-moisture anomalies) cannot be interpreted with the classical recycling mechanism". Seneviratne et al. (2010) summarise that "the key for understanding soil moisture-precipitation interactions lies more in the impact of soil moisture anomalies on boundary-layer stability and precipitation formation than in the absolute moisture input resulting from modified evapotranspiration. For instance, the additional precipitated water falling over wet soils may originate from oceanic sources, but the triggering of precipitation may itself be the result of enhanced instability induced by the wet soil conditions". However, the cited studies focus on subcontinental scales. Moisture recycling may still be important for continental-scale atmospheric moisture transport because the land-atmosphere moisture exchange remains being subject to mass conservation. Even if an important aspect for understanding evaporation-precipitation interactions lies in the local to regional interactions, traditional moisture recycling may have its place in the large-scale picture.

The land-atmosphere coupling studies cited above focus on the active role of moisture in the atmosphere. In contrast, in the traditional moisture recycling studies moisture is treated as if it were a passive atmospheric tracer. This is particularly obvious where precipitation is formulated as a function of the atmospheric moisture content (and horizontal windspeed) only, as for example in Wiesner (1970) and Savenije (1995b). With this assumption, i.e. without local interactions, a reduction of land-surface evaporation would lead to a proportional moisture loss of the passing atmospheric air. The main question is to what extent local interactions change the situation through their influence on local precipitation. In principle the full range of precipitation-responses is possible: in case of a negative evaporation-precipitation coupling the moisture depletion is amplified. In case of no coupling the moisture depletion goes in line with the traditional moisture recycling view. Finally, in case of a positive evaporation-precipitation coupling, which seems to be the rule rather than the exception (Seneviratne et al., 2010), a local precipitation reduction may partly or fully compensate, or even overcompensate for the missing evaporation. In case of substantial compensation the atmospheric moisture content would remain largely unaffected by the evaporation reduction. Would recycling estimates still be able to tell something about downstream consequences of upstream land-cover change?

To explore the meaning of continental moisture recycling estimates we conduct two equilibrium experiments with an atmospheric general circulation model with strongly differing land-surface properties: while in the first experiment the land-surface parameterisation resembles present-day conditions, continental evaporation is completely suppressed in the second experiment (Sect. 2.1). We use the modelled fields of moisture, wind, and evaporation to compute spatio-temporally resolved continental RMFs for the present-day experiment with a vertically integrating tracing scheme as done in Yoshimura et al. (2004) and van der Ent et al. (2010) (Sect. 2.2). Since continental RMFs are zero per definition in the second experiment, we can compare the present-day RMF estimates (Sect. 3) directly with the modelled hydrological response following the suppression of continental evaporation (Sect. 4). After a schematic illustration of

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our results (Sect. 5) we discuss the limitations of our study (Sect. 6) and draw conclusions (Sect. 7).

2 Methods

2.1 Model experiments

For our investigations we use the Earthsystem model of the Max Planck Institute for Meteorology (MPI-ESM), comprising the atmospheric general circulation model ECHAM6 (Roeckner et al., 2003), including the land-surface scheme JSBACH (Raddatz et al., 2007), at T63/L47 resolution (1.875° × 1.875°, 47 levels, 10 min time step). We do not use the interactive MPI-ESM ocean component, but prescribe climatological sea-surface temperatures (SSTs) representing present-day conditions without interannual variability.

We run the model in two configurations. The reference experiment "REF" represents present-day conditions. In the second experiment "DRY" the continents are not allowed to exchange moisture with the atmosphere through evaporation (and, less importantly, condensation) but only through precipitation. Thereby the continents behave essentially as if they were kept completely dry. In reality a similar hydrological behaviour of the continents could in principle be provoked by transforming the continents into coarse-textured or rocky deserts with sufficiently steep slopes allowing for exhaustive runoff.

Since we focus on the direct effect of continental evaporation, or rather its absence, we prescribe continental albedo and roughness from climatologies in both experiments. This reduces the number and complexity of interactions and feedbacks that would otherwise add secondary alterations to the modelled differences in climate. The albedo and roughness climatologies stem from a multi-year equilibrium model run with dynamically modelled albedo and roughness that is otherwise identical to the REF experiment. The two experiments REF and DRY span 34 years each, but we exclude the first four years from further analyses.

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2.2 Moisture tracing

We destinguish two types of atmospheric moisture: oceanic and recycled. While oceanic moisture stems from ocean evaporation, recycled moisture stems from continental evaporation. The recycled moisture fraction (RMF) is the fraction of recycled moisture in total moisture.

$$\text{RMF} = \frac{M_r}{M_r + M_o} \tag{2}$$

where M_r is recycled moisture and M_o is oceanic moisture. In principle this measure is defined at every point in space and time in the atmosphere for infinitesimal volumes. We, however, apply the well-mixed assumption which implies that the RMF is assumed to be vertically constant. Accordingly, M_r and M_o are vertically integrated moisture densities. Ignoring temporarily that horizontal wind velocities generally vary with height, the problem reduces to two spatial dimensions. With these simplifications we can consider a Lagrangian atmospheric column travelling horizontally with the mean wind. The source and sink terms for M_r and M_o now read

$$\frac{dM_r}{dt} = E_r - \text{RMF} \cdot P \tag{3}$$

$$\frac{dM_o}{dt} = E_o - (1 - \text{RMF}) \cdot P \tag{4}$$

where t is time, P is precipitation from the air column, and E_r is evaporation into the air column from land, and E_o is evaporation into the air column from the ocean. Using Eqs. (2–4), substituting $M_r + M_o$ by VIM (vertically integrated moisture density), and transforming into Eulerian formulation yields

$$\frac{\partial \text{RMF}}{\partial t} + u_{\text{eff}} \frac{\partial \text{RMF}}{\partial x} + v_{\text{eff}} \frac{\partial \text{RMF}}{\partial y}$$
$$= \frac{E_r}{\text{VIM}} \cdot (1 - \text{RMF}) - \frac{E_o}{\text{VIM}} \cdot \text{RMF} \tag{5}$$

where u_{eff} is the effective wind component along the zonal coordinate x, and v_{eff} is the effective wind component along the meridional coordinate y. Since the air column can not be over land and over the ocean at the same time, only one of the right-hand-side terms can be nonzero at a time, depending on the location. The effective wind components in this 2-dimensional formulation are vertical mean values weighted by the local water vapour partial pressure, such that

$$
\begin{aligned}
F_u = u_{eff} \cdot \text{VIM} &= u_{eff} \cdot \left(g^{-1} \cdot \int_0^{p_0} q(p)\, dp \right) \\
&= g^{-1} \cdot \int_0^{p_0} u(p) \cdot q(p)\, dp
\end{aligned}
\tag{6}
$$

where F_u is the zonal component of the vertically integrated horizontal moisture flux, g is gravitational acceleration, p is pressure, p_0 is surface pressure, and q is specific moisture (all phases). Equation (6) analogously applies to the meridional component of the vertically integrated horizontal moisture flux, F_v. Equations (5) and (6) can also be derived from vertical integration of the full 3-dimensional equations (not shown).

Equation (5) reveals that precipitation, although occuring in Eqs. (3) and (4), does not influence the RMF directly. The reason is that, with the well-mixed assumption, precipitation removes recycled moisture and oceanic moisture from the atmospheric column in proportion to their respective abundance. Precipitation affects the RMF only indirectly through its effect on vertically integrated moisture.

We discretise Eq. (5) with upstream differencing (Press et al., 2007) and run the algorithm over all 34 model years of the REF experiment, but incorporate only the last 30 yr into our analysis. Since there is no continental moisture in the DRY experiment, there is no need to apply the tracing to it.

3 Recycling patterns

The relative abundance of recycled moisture in the atmosphere (Fig. 1, left) is determined by the rate of surface evaporation, the horizontal moisture flux density, and the land-sea geometry (Fig. 1, right). As a result the RMF increases from continental upstream coasts to downstream coasts with respect to the prevailing winds. Steep RMF gradients occur where strong evaporation combines with moderate horizontal moisture-flux density (e.g. tropical Africa), or where the air flows perpendicular to a steep evaporation gradient (e.g. Sahel, particularly in January), or a combination thereof (e.g. China in July).

Under present-day conditions recycled moisture contributes up to 80% to the atmosphere's total water content. This peak value is reached in Central Asia, Siberia, and the north-eastern parts of North America in July. At this time of the year the RMF exceeds 60% in the whole Arctic region, meaning that the main sources of moisture are the large land masses enclosing the pole. In January the RMF in the northern extratropics hardly reaches 20% because of strongly reduced land-surface evaporation. The RMF in the southern extratropics is low even during the Southern-Hemisphere summer (January) because of the absence of comparably large land masses.

In the tropics the RMF peaks at about 50% in the downstream regions of Africa and South America with weak seasonality. Despite strong land-surface evaporation, recycled moisture does not accumulate in the tropics as much as it does in the northern extratropics during summer, because air travelling between the tropical continents encounters large, strongly evaporating ocean basins. In consequence, air reaching South America or Africa contains almost no recycled moisture. This is different in the northern extratropics, where the RMF is still relatively high after an ocean crossing. Additionally, horizontal moisture flux densities are usually higher in the tropics because the warmer atmosphere contains more moisture (Fig. 1, right).

Our results agree well with RMF estimates published by Numaguti (1999), Bosilovich et al. (2002), Yoshimura et al. (2004), and van der Ent et al. (2010). The similarity of

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the estimates obtained with 2-dimensional tracing, which our study has in common with Yoshimura et al. (2004) and van der Ent et al. (2010), compared to the estimates obtained with 3-dimensional tracing (Numaguti, 1999 and Bosilovich et al., 2002), suggests that the error introduced by the vertical integration is acceptably small for our
5 large-scale considerations.

4   Response to the suppression of continental evaporation

4.1   Response of the atmospheric moisture content

Based on traditional moisture-budget considerations one would expect that the response of the atmospheric moisture content to the suppression of continental evap-
10 oration is converse to the RMF pattern in the REF experiment (Fig. 1, left). But this is not the case (Fig. 2). Strikingly, the patterns show hardly any correlation. The discrepancy is most evident in July, when the atmosphere in the Arctic carries 20% more water in the DRY experiment, although around 60% of the atmospheric water in the REF experiment stem from continental evaporation. Furthermore, VIM does not drop
15 systematically from continental upstream to downstream regions. The atmosphere is substantially drier by up to 50% in a large region ranging from the Sahel to central Asia, while at the center of this region, namely in northern Africa, the RMF has a local minimum. The western half of Eurasia is much more affected by decreased atmospheric moisture than the eastern half, although the opposite is expected based on the RMF
20 pattern. Around the eastern coast of Asia the atmosphere carries even 10–15% more moisture with suppressed continental evaporation.

The discrepancy between the RMF and the change in VIM is similarly apparent not only in North America, but also in the tropics. In tropical Africa during July the RMF peaks north of the equator, in particular in tropical West Africa. In contrast, the atmospheric moisture content is unchanged in tropical West Africa, but considerably decreased south of the equator along Africa's western coast. In January, when the patterns agree comparably well in tropical Africa, VIM is almost unchanged in most of South America but decreased in its southern part, whereas high RMFs occur in the central part. Although it seems that the changes in the atmospheric moisture content can partly be explained with the RMF patterns in some regions, for example around India and China during January, the overall contradiction casts doubt on a general causal relation.

The picture becomes more complex when the vertical distribution of humidity changes is taken into account (Fig. 3). While the lower troposhere is consistently drier in the DRY experiment (despite the temperature increase that in principle allows for more moisture, see Sect. 4.3), the middle atmosphere is moister. This distinct vertical structure indicates that the dry-anomaly at the surface is not rapidly mixed into higher altitudes through the stably stratified free atmosphere. Rather, higher atmospheric layers over the continents are influenced by both the drier continental boundary layer and the oceanic boundary layer, the latter being moister particularly in high northern latitudes (not shown). Irrespective of its causes, this distinct vertical structure indicates that results obtained with vertically integrative moisture-budget models should be taken with a grain of salt.

4.2 Response of precipitation

The response of precipitation is clearly different over the ocean compared to the continents, in particular in July: while the continents receive much less precipitation in the DRY experiment the response over the ocean is ambiguous (Fig. 4). This may not astonish at first glance recalling that the introduced difference between the experiments concerns the land-surface only. However, taking the traditional moisture-budget

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approach serious, the hydrological response should follow the RMF-pattern (Fig. 1, left) over the ocean as over land: the more a region is supplied by continental evaporation in the REF experiment, the more of a drying is expected to affect the region in response to the suppression of continental evaporation. But this is not the case, most obviously in the northern extratropics in July: while over the ocean the fraction of moisture in the atmosphere that stems from continental evaporation (∼60%) is almost as large as over land (∼60–80%), only the land is affected by severe precipitation loss.

The ambiguous response of precipitation over the ocean indicates that the hydrological state of the atmosphere has a short memory over the ocean, meaning that the hydrological state returns to a quasi-equilibrium humidity state much faster than it takes the air to travel continental distances. This finding, which is also in line with the response of the atmospheric moisture content (Sect. 4.1), is not surprising given the fact that for atmospheric considerations the ocean is an inexhaustible water reservoir, and that the boundary layer, which contains most of the atmospheric moisture, is generally well-mixed. This suggests that continental moisture recycling can not act across large ocean basins, i.e. inter-continental, but only intra-continental. To give a simple example, Eurasia is not affected by North America's evaporation and vice versa, regardless of the substantial fraction of moisture they receive from each other. Over land the situation may be different: since evaporation is constrained by precipitation the surface can dry up and, consequently, atmospheric dry-anomalies can persist. To account for the evidence that continental moisture recycling can not act across large ocean basins, in the following we put our focus on intra-continental gradients of the RMF.

Continental precipitation rates respond strongly to the suppression of continental evaporation (Fig. 4). In the northern extratropics continental precipitation is almost completely absent during July – in upstream regions as in downstream regions. Europe and the western parts of North America are as much affected as the eastern parts of the continents. In July, precipitation in southern Africa, which is already low under present-day conditions, decreases by almost 100%, although the RMF indicates that

under present-day conditions only about 10% of the atmospheric moisture is of continental origin. The situation is similar in Australia. Also in January it is hard to find meaningful correlations between the RMF pattern and the precipitation response. As it is the case with the atmospheric moisture content, the mismatch between the RMF and the precipitation response casts doubt on a stringent causal relation between the two.

To understand the consequences of the precipitation response for the atmospheric large-scale moisture transport, it is helpful to analyse how the aerial runoff (precipitation minus evaporation, $P - E$) changes between the experiments. If precipitation did not respond to the suppression of continental evaporation, the land would become a stronger moisture sink exactly by the present-day evaporation rate (Fig. 1, right). Within the traditional moisture recycling framework some precipitation decrease is expected following the RMF pattern, which would alleviate the amplification of the continental moisture sink. This alleviating (compensating) effect would occur in continental downstream regions (with low RMF) more than in upstream regions (with high RMF). However, since precipitation responds strongly not only in continental downstream regions, but also in upstream regions, the precipitation decrease compensates for most of the evaporation decrease in large parts of the continents (Fig. 5, left). Paradoxically, the suppressed continental evaporation flux is often fully or even over-compensated, the latter meaning that the suppression of evaporation does not strengthen but weaken the land moisture sink. This happens in large parts of Eurasia and North America, in particular during northern summer, and in the tropics around the Intertropical Convergence Zone (ITCZ).

Where the land is a moisture source ($P - E < 0$) for some time of the year in the reference situation (Fig. 5, right), the precipitation response can not fully compensate for the missing evaporation, even if precipitation rates become zero. Strikingly, regions with weak compensation are located almost exclusively where $P - E$ is negative in the reference situation for some time of the year, foremost in tropical wet-dry climates during the dry season when the ITCZ is located on the other side of the equator, and in

substantial parts of the northern extratropics in July. Where precipitation exceeds evaporation in the REF experiment, the suppressed evaporation is largely compensated by the precipitation response. In consequence the continents do not become extensive moisture sinks. This in turn explains why in general there is no negative VIM-anomaly accumulating from upstream to downstream continental regions (Fig. 2).

On the other hand this reasoning implies that in those regions mentioned above, where evaporation exceeds precipitation for some time of the year, moisture recycling in the traditional sense is inevitably taking place to some extent. As expected from moisture-budget considerations, in these regions decreased downstream-precipitation is accompanied by a negative VIM-anomaly accumulating from upstream to downstream. It therefore seems that precipitation particularly in (I) South America around the Tropic of Capricorn in July, (II) the western parts of southern Africa in July, and (III) China in January relies at least partly on continental upstream evaporation. The corresponding upstream moisture-source regions are (I) Amazonia, (II) tropical southern Africa, and (III) India and Indochina. Hence, the reduction of downstream precipitation in these regions can at least partly be ascribed to traditional moisture recycling, though not completely because in our experimental setup the downstream regions are also directly affected by changed land-surface conditions combined with local to regional evaporation-precipitation coupling.

4.3 Response of temperature

To understand why the correlation between the RMF patterns in the REF experiment and the modelled hydrological changes is mostly poor, we now consider the response of other climatic key variables, first of all temperature. Without evaporative cooling the continents are substantially warmer (Fig. 6). In July near-surface temperatures are higher by up to 16 K in central Asia, and by almost as much in eastern Europe and central North America. While the northern extratropical warming is almost neglegible in January, because there is already not much evaporation in the REF experiment, the large tropical continents are warmer by up to 9 K throughout the year. Overall, the

continental near-surface warming closely follows the evaporation rates under present-day conditions (Fig. 1, right). The northern extratropical warming in July is additionally amplified by strongly decreased cloud cover (not shown).

The warming is not confined to the surface, but stretches high into the atmosphere (Fig. 7). In the tropics the continental warming converts into a slight cooling at 500–600 hPa because reduced deep convection and precipitation is associated with reduced condensation-induced heat release in the atmosphere. However, the northern extratropical warming in July extends up to the tropopause where it still amounts to 2 K. One reason for this difference between the tropics and the extratropics regarding the vertical temperature-change profile is the decreased extratropical cloud cover, which contributes to overcompensate the high-altitude cooling tendency. Also, in the ITCZ, where the spatially focussed high $P - E$ is fed by a large area with negative $P - E$ including the subtropics, the condensation-induced heating term is much more important compared to other diabatic terms than it is the case in the extratropics. Finally, the hydrological cycle is not just redistributing heat from the surface to the atmosphere, leaving the overall heat budget unchanged, but causes an overall cooling of the system: the longwave radiation emitted from higher altitudes escapes more readily into space. The weaker global hydrological cycle in our DRY experiment is therefore not associated with an extensive cooling in the free troposphere.

The strong continental warming provides an explanation for the marked mismatch between the RMF patterns in the REF experiment and the modelled hydrological changes. For a given total moisture content, higher atmospheric temperatures result in decreased relative humidity. This tendency is amplified by strongly reduced specific humidity in the lower continental atmosphere (Fig. 3): the dewpoint spread increases by as much as 50 K in the continental boundary layer around 50° N on zonal average (not shown). This in turn suppresses cloud formation, moist convection, and precipitation. Decreased precipitation rates compensate for the missing moisture input from evaporation, such that the moisture content of an atmospheric column travelling over a continent may evolve similarly both with and without suppressed continental evaporation.

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While traditional moisture-budget considerations suggest that the suppression of continental evaporation results in progressive drying from upstream to downstream regions, the temperature response changes the situation markedly. Keep in mind, however, that the situation is different in regions with negative $P - E$ (see Sect. 4.2).

Some care has to be taken regarding the overall effect of the temperature increase because the differences are vertically not homogeneous. Since the surface warms stronger than the higher troposphere the average stability of the atmospheric stratification weakens, which favours the occurence of convective precipitation. This may counteract the effect of warming-caused decreased relative humidity to some extent, but according to our results the latter by far dominates the total effect.

4.4 Response of the atmospheric circulation

Atmospheric circulation patterns do not remain unaffected by the changes following the suppression of continental evaporation. The extensive summer-warming of the continents and the strengthening of the associated thermal lows result in stronger monsoonal circulations. In particular the South Asian Monsoon is amplified, resulting in increased precipitation rates in northern India in July (Fig. 4). The effects of changes in the large-scale circulation are not constrained to the continents, but act on regions as remote as the central Pacific Ocean. Both here and in the Indian Ocean a northward shift of the ITCZ results in dipol-like precipitation differences in July. At the same time a southward shift of the ITCZ over Africa acts to keep precipitation rates relatively high around 5° N while the surroundings experience severe drying. The pronounced changes in remote oceanic regions are particularly astonishing given the fact that the sea-surface temperatures are fixed between the experiments, showing that the atmospheric large-scale dynamics are broadly influenced by continental conditions.

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5 A schematic illustration of our results

Figure 8 illustrates schematically the basic differences between the "traditional" and the "interactive" view on a hypothetic atmospheric response following the suppression of continental evaporation. The figure shows profiles of the atmospheric moisture content and surface fluxes as 1D-transects through an idealised continent along the prevailing wind direction, with oceans situated upstream and downstream. The traditional case is the response one might expect from pure moisture-budget considerations. The interactive case is an idealised response of a dynamically reacting atmosphere.

Before landfall the air is in a quasi-equilibrium humidity state over the ocean, meaning that precipitation and evaporation are in balance ($P - E = 0$). In the reference situation (Fig. 8, left), the evaporative fraction over the continent is assumed to be constant at 60%, meaning that 40% of the precipitation are removed from the system as runoff. With precipitation assumed to be a function of VIM only, here exemplarily as $P \propto \text{VIM}$ (compare Wiesner, 1970; Savenije, 1995a), VIM and the surface fluxes decrease continuously (exponentially) from the upstream coast to the downstream coast. At the same time the recycled moisture fraction increases. When the air leaves the continent it returns to its initial equilibrium, with the ocean providing the moisture needed to remove the deficit ($P - E < 0$).

Like in the reference situation, in the traditional case (Fig. 8, middle) precipitation is assumed to be a function of the atmospheric moisture content only ($P \propto \text{VIM}$) and evolves continuously with the atmospheric moisture content. Without continental evaporation the progressive atmospheric moisture loss is stronger than in the reference situation. The relative reduction of both VIM and precipitation compared to the reference situation (bottom graph) increases from upstream to downstream – in parallel with the increase of the recycled moisture fraction in the reference situation.

In the interactive case (Fig. 8, right) precipitation is still a function of VIM, but additionally responds to changes in other atmospheric variables induced by the suppression of evaporation. Motivated by our results we assume that precipitation decreases sharply

at the point of landfall due to the continental temperature increase, which in turn results from the absence of surface evaporation. Instead of making up the details of the functional dependencies, we show the special case that the precipitation decrease exactly compensates for the suppressed evaporation. Indeed this gross simplification is not too far from our modelling results (Fig. 5, left). In consequence, the atmospheric moisture content evolves as in the reference situation, and the relative precipitation decrease is 60 % on the whole continent – the traditionally expected upstream to downstream drying-gradient vanishes (bottom graph).

The idealised profiles shown in Fig. 8 are overly simplistic: the real atmosphere is not 1-dimensional but 3-dimensional, large-scale wind fields are not constant but vary in time, most importantly on synoptic and seasonal scales, and precipitation is a complex function of the atmospheric state. Also, as shown in Sect. 4.4, the atmospheric circulation is affected by the suppression of continental evaporation. Climatic seasonality makes it possible that some continental regions are significant moisture sources for some time of the year, which is another complication ignored in this time-invariant picture. In such circumstances the precipitation reduction can not completely compensate for the missing evaporation, as assumed in the idealised interactive case, which therefore can be only partly valid for regions with considerably negative $P - E$ in the reference situation (Fig. 5, right). On the other hand there are large regions where the precipitation reduction even overcompensates for the missing evaporation (Fig. 5, left). The most important aspect illustrated in Fig. 8, however, is the upstream to downstream gradient of the continental drying, which vanishes when the full response of the atmosphere is taken into account. In this aspect the interactive view fits better to the experimental results than the traditional view.

6   Discussion

The quality of our results depends on (I) the reliability of the Earthsystem model, (II) the reliability of the moisture tracing, and (III) the adequacy of the experimental setup.

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The MPI-ESM is among the most widely used global climate models and is capable of reproducing the most important aspects of present-day climate (e.g. Hagemann et al., 2006). This does not guarantee that the model behaves realistically under strongly perturbed conditions, as it is the case in our DRY experiment. It has
5 been shown that the spread between different atmospheric models regarding local soil moisture-precipitation coupling is quite large (Koster et al., 2002, 2006). However, this finding is based on GLACE-type experiments (GLACE standing for Global Land-Atmosphere Coupling Experiment) which focus on naturally occuring inter-annual soil moisture anomalies rather than on large-scale land-cover change, which is an impor-
10 tant difference to our study. Modelling studies investigating the climatic response to extreme land-cover change scenarios, for example global removal of vegetation (Betts, 1999; Kleidon et al., 2000; Fraedrich et al., 2005), indicate that the inter-model spread is smaller when it comes to extreme land-cover change. The pioneering study by Shukla and Mintz (1982) includes a model experiment that is similar to our DRY ex-
15 periment, and they find a similar climate response. It therefore seems that the model results are not as uncertain as one might expect based on GLACE-type experiments. Nevertheless our findings claim for validation through similar studies using other models.

The computed recycling patterns are based on a vertically integrating tracing scheme
20 (Sect. 2.2). It is evident that this simplification introduces some error into the estimates, in particular where the usually stably stratified free atmosphere is seldomly mixed in the vertical through high convection and/or where horizontal winds are substantially sheared in the vertical. The latter is typically not so much the case in the extratropics, where the flow is dominated by cyclones and anticyclones, but may be more relevant
25 closer to the equator. For example in tropical western Africa during northern summer a north-east directed moisture flux in the monsoon layer below 750 hPa is essentially compensated by a reverse flow above 750 hPa. The weak horizontal moisture flux remaining after vertical integration presumably results in overestimation of moisture recycling. However, it seems that the errors are acceptably small given the similarity

of recycling estimates obtained with (Yoshimura et al., 2004; van der Ent et al., 2010, and this study) and without (Numaguti, 1999; Bosilovich et al., 2002) vertical integration. Furthermore, the focus of our study is on continental-scale patterns, in particular upstream to downstream gradients, rather than on accurate regional estimates.

⁵ Our experimental setup is designed in such a way that continental moisture recycling estimates can be compared directly to the actual climatic response following the total suppression of the source of recycled moisture, namely continental evaporation. This direct comparability is achieved at the expense of realism: the complete suppression of any continental evaporation is far from any realistic land-cover change scenario. We

¹⁰ can not rule out that recycling estimates gain significance to infer precipitation changes when the land-cover modifications are more realistic, i.e. less extreme in spatial extent and in the degree of evaporation reduction. While realistic scenarios are accompanied by weaker changes of the large-scale circulation, we expect that the effect of local to regional evaporation-precipitation coupling still impairs the significance of recycling

¹⁵ estimates, perhaps to the same degree as with our large-scale experimental setup. Accordingly targeted modelling studies could help to improve the current understanding of the issue of scale-dependence of evaporation-precipitation coupling and moisture recycling.

Regarding the issue of scale-dependence, we stress that the coupling strengths

²⁰ shown in Fig. 5 (left) must not be interpreted as the result from local evaporation-precipitation coupling alone. The local precipitation response is the result of local, regional, and large-scale effects attributable to the applied large-scale land-cover change. An interesting question that we do not further investigate here is how strongly the patterns would deviate from those shown in Fig. 5 (left) if the suppression of evap-

²⁵ oration was applied to the model grid-cells one by one, or to regions of some intermediate spatial extent.

To keep the interpretability of our results as clear as possible, we use identical climatologies of surface albedo, surface roughness, and SSTs in both experiments. The climatic response of the fully dynamical atmosphere-ocean-land system to the

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suppression of continental evaporation could be considerably different from the response we obtain with our stronger controlled setup. Important missing feedbacks involve for example albedo changes associated with the extent of snow and sea-ice covered areas. However, our setup is not designed to yield a most realistic full-dynamics response, but to assess the meaning of moisture recycling estimates. We have no reason to assume that one or more of the missing feedbacks would change the response in such a way that our conclusions regarding moisture recycling would have to be revised significantly.

7 Summary and Conclusions

The suppression of continental evaporation strongly alters the Earth's climate. The severe reduction of continental precipitation, and in particular its spatial pattern, can not be explained by simple atmospheric moisture-budget considerations as implied by the traditional concept of moisture recycling. The continents' property to act as a sink for atmospheric moisture is not substantially amplified, as one might expect intuitively. Instead, reduced precipitation rates due to strong positive local to regional evaporation-precipitation coupling compensate for most of the missing evaporation. Hence, the atmospheric moisture content is not systematically reduced. The evaporation-precipitation coupling is associated with a strong increase of continental temperatures due to the absence of evaporative cooling at the surface. While accordingly the warming is strongest at the surface, it reaches far into the free atmosphere, in particular in the extratropics. Most importantly, the warming and the associated precipitation decrease is not confined to continental downstream regions, but occurs in upstream regions as well.

From our results it seems that moisture recycling estimates are of limited use to deduce hydrological impacts of land-cover change activities. Over the last two decades evidence has grown that hydrological land-surface properties influence climate locally to regionally via evaporation-precipitation coupling rather than remotely

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via the atmospheric moisture-budget. On the other hand, it still seemed reasonable that moisture-budget considerations are relevant at least at continental scales, because after all the water in the atmosphere-land system remains being subject to mass conservation. We find evidence that this is the case in continental regions that
5 are substantial moisture sources for some time of the year, foremost tropical wet-dry climates during the dry season. Apart from these exceptions, our results question the relevance of traditional moisture recycling estimates even for continental scales – an admittedly counterintuitive conclusion. More generally, our results imply that source-target relations of water are of limited use to infer potential impacts at the target
10 due to changes at the source, as would be the case for a truly passive atmospheric tracer.

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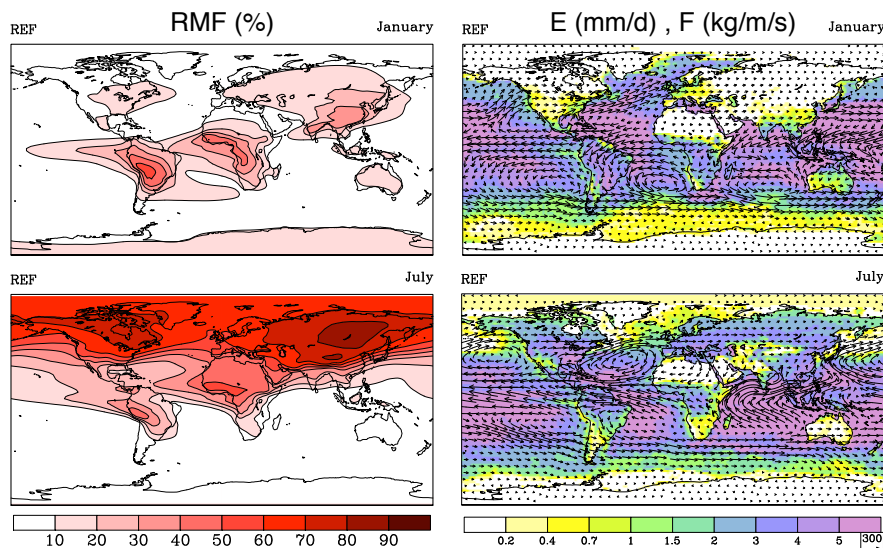
Fig. 1. Left: fraction of recycled moisture (RMF, %) in the REF experiment. Right: surface evaporation (colours, $mm\,d^{-1}$) and horizontal moisture flux density (arrows, $kg\,m^{-1}\,s^{-1}$) in the REF experiment. The reference arrow in the lower right corner defines the scale of the moisture flux density.
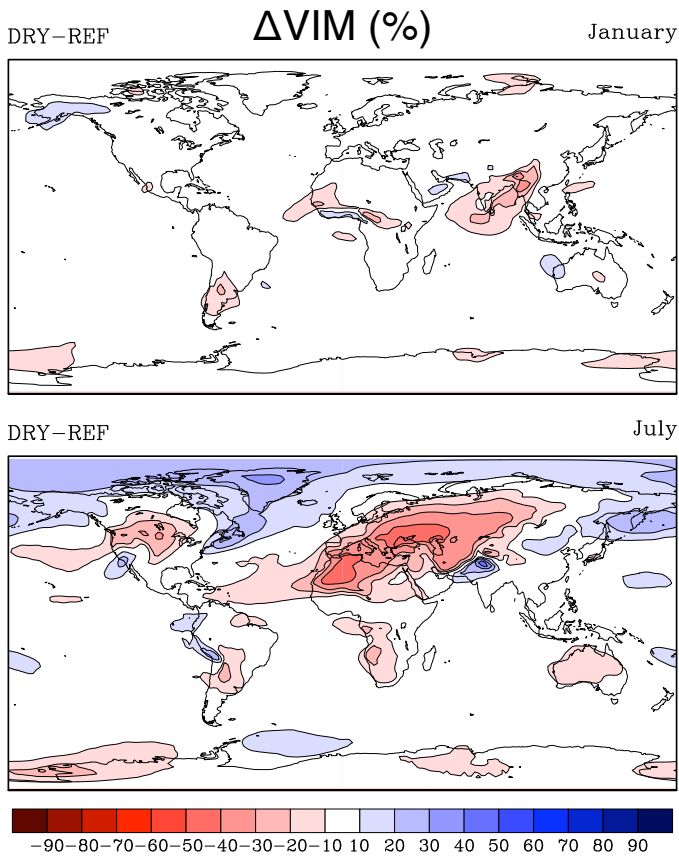
Fig. 2. Relative VIM difference ($\frac{DRY-REF}{\max(DRY,REF)}$, vapour + liquid + ice, %). Note that the red part of the colour scale is kept identical to the one used in Fig. 1, left, to allow for direct comparison.
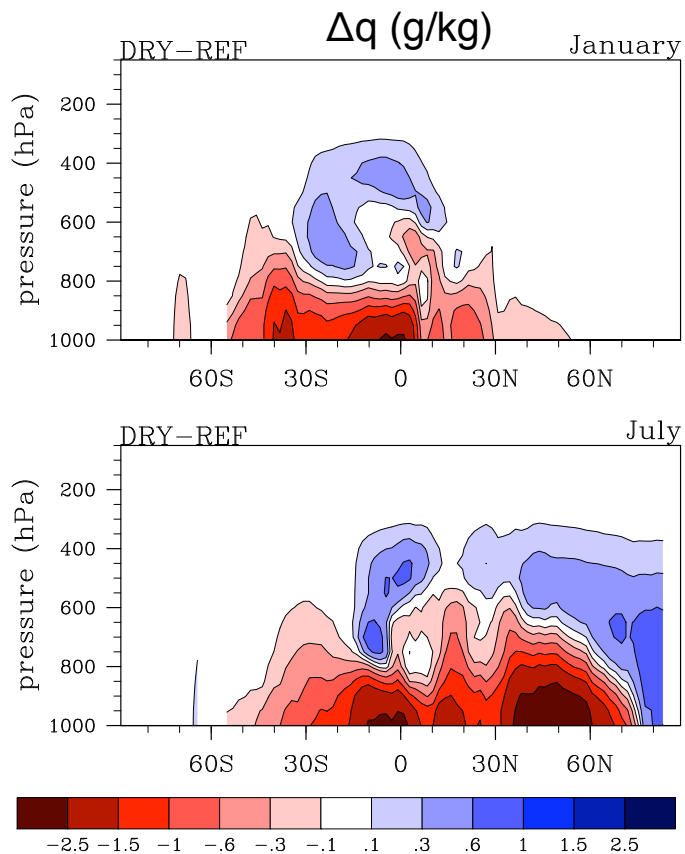
Fig. 3. Difference in zonal-mean specific humidity (g kg^{-1}), averaged over land only.

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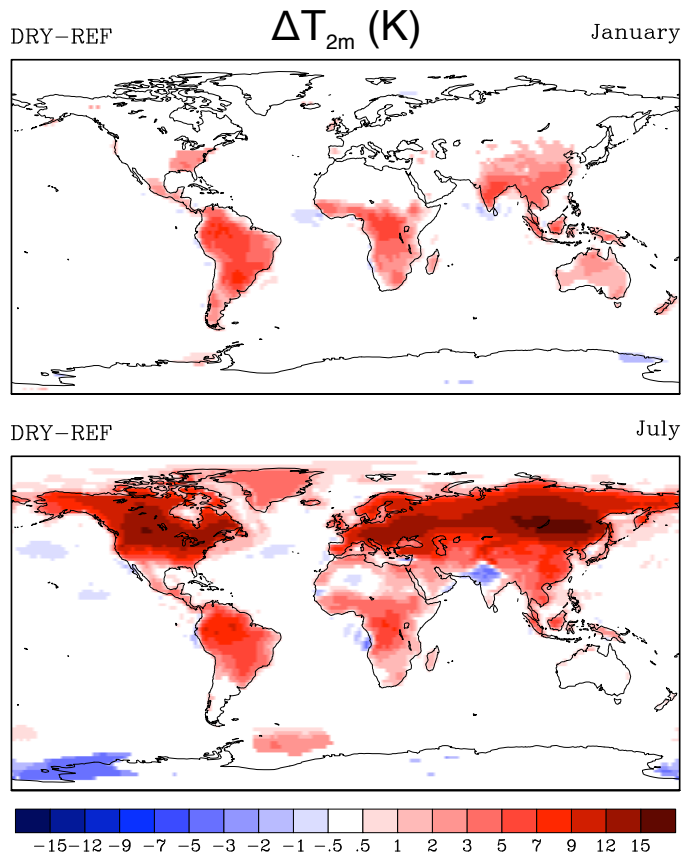Full Screen / Esc

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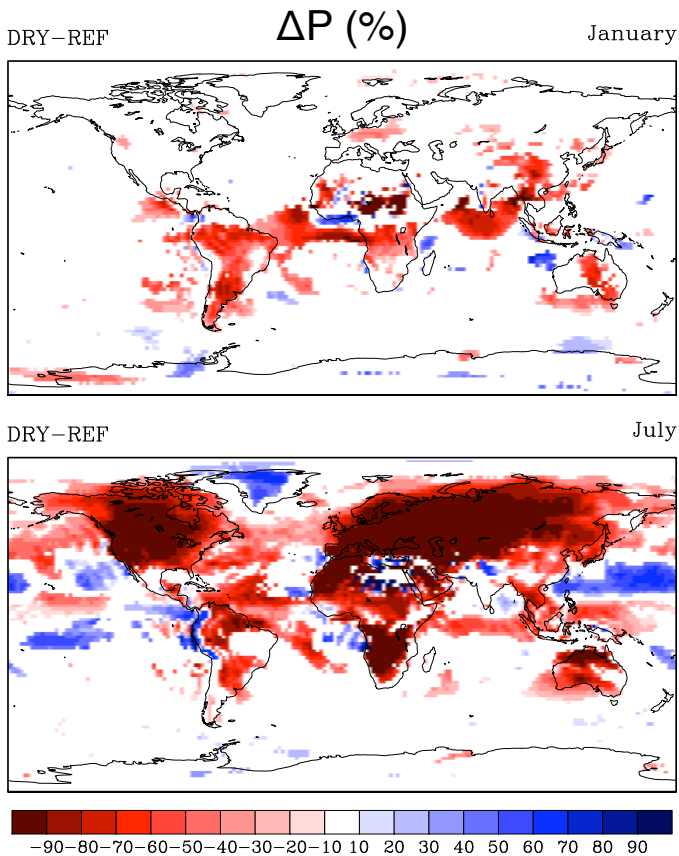
Fig. 4. Relative precipitation difference ($\frac{\text{DRY}-\text{REF}}{\max(\text{DRY},\text{REF})}$, %). Only differences significant at the 99%-level are shown (Wilcoxon rank-sum test, no significant 1-year lag autocorrelations in the data). Note that the colour scale is kept identical to the one used in Fig. 1, left, and Fig. 2 to allow for direct comparison.
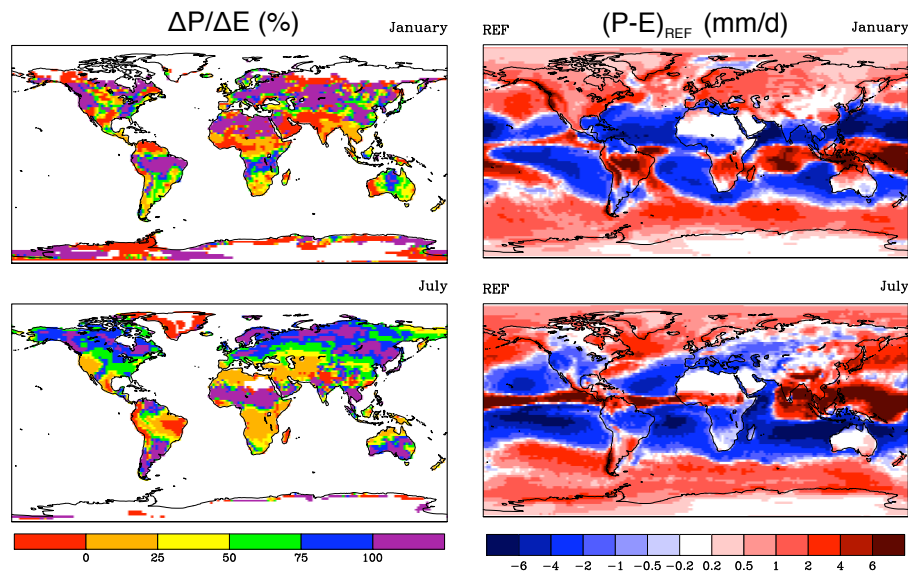
Fig. 5. Left: compensation of the missing continental evaporation through reduced precipitation ($\frac{P_{REF}-P_{DRY}}{E_{REF}}$, %). Violet indicates over-compensation (the land becomes a weaker moisture sink) and hence points at a very strong positive evaporation-precipitation coupling. Blue, green, yellow and orange indicate strong to weak positive coupling, and red indicates a precipitation increase, i.e. a negative evaporation-precipitation coupling. Continental regions with negative evaporation (= dew) in the REF experiment are left white. Note that the coupling strengths shown here must not be interpreted as the result from local evaporation-precipitation coupling alone, see Sect. 6. Right: aerial runoff ($P - E$, mm d^{-1}) in the REF experiment. Positive (negative) values indicate that the surface is a moisture sink (source).

Fig. 6. Near-surface (2m) temperature difference (K). Only differences significant at the 99%-level are shown (Wilcoxon rank-sum test, no significant 1-year lag autocorrelations in the data).
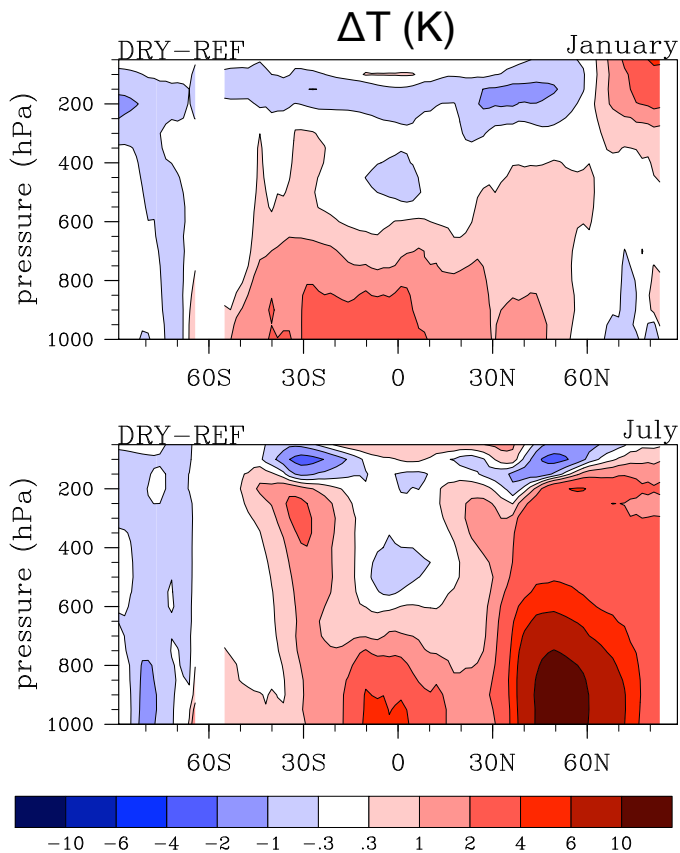
Fig. 7. Difference in zonal-mean temperature (K), averaged over land only.

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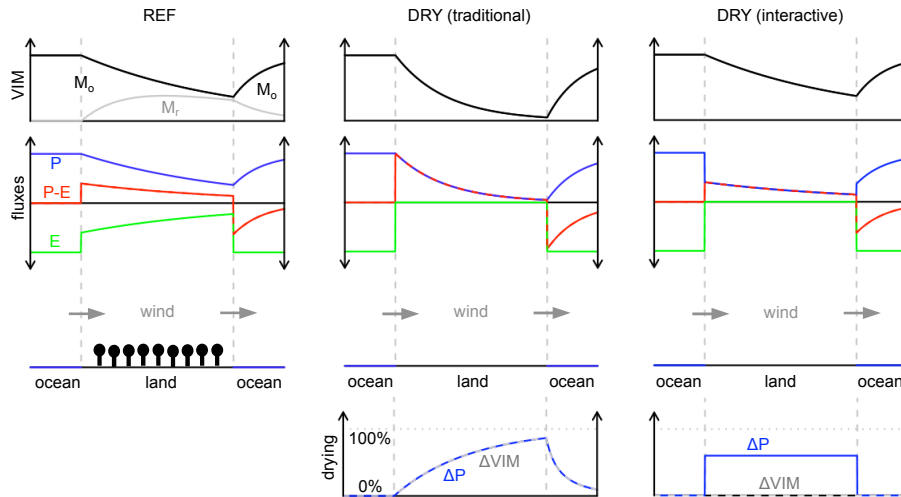
Fig. 8. Illustration of the qualitative difference between the traditional view (without local interactions/coupling) and the interactive view (with local interactions/coupling) as 1D-transects through an idealised continent. Left: reference situation with continental evaporation (as in the REF experiment). Middle and right: expected response to the total suppression of continental evaporation (as in the DRY experiment) according to the traditional view and according to an idealised interactive view. VIM = vertically integrated moisture, M_o = oceanic moisture, M_r = recycled moisture, P = precipitation, E = evaporation. Bottom graphs show the relative reduction ($\frac{\text{REF}-\text{DRY}}{\text{REF}}$) of P (blue, solid) and VIM (grey, dashed). Units are arbitrary.

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