Contrasting roles of interception and transpiration in the hydrological cycle. Part 1: temporal characteristics over land

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Abstract. Moisture recycling, the contribution of terrestrial evaporation to precipitation, has important implications for both water and land management. Although terrestrial evaporation consists of different fluxes (i.e., transpiration, vegetation interception, floor interception, soil moisture evaporation, and open water evaporation), moisture recycling (terrestrial evaporation-precipitation feedback) studies have up to now only analysed their combined total. This paper constitutes the first of two companion papers that investigate the characteristics and roles of different evaporation fluxes for land-atmosphere interactions. Here, we investigate the temporal characteristics of partitioned evaporation on land, and present STEAM (Simple Terrestrial Evaporation to Atmosphere Model) – a hydrological land surface model developed to provide inputs to moisture tracking. STEAM estimates a mean global terrestrial evaporation of 73 900 km$^3$ year$^{-1}$, of which 59% is transpiration. Despite a relatively simple model structure, validation shows that STEAM produces realistic evaporative partitioning and hydrological fluxes that compare well with other global estimates over different locations, seasons and land-use types. Using STEAM output, we show that the terrestrial residence time scale of transpiration (days to months) has larger interseasonal variation and is substantially longer than that of interception (hours). Most transpiration occurs several hours or days after a rain event, whereas interception is immediate. In agreement with previous research, our simulations suggest that the vegetation’s ability to transpire by retaining and accessing soil moisture at greater depth is critical for sustained evaporation during the dry season. We conclude that the differences in temporal characteristics between evaporation fluxes are substantial and reasonably can cause differences in moisture recycling, which is investigated more in Part 2, the companion paper.

1 Introduction

Terrestrial evaporation is mediated by land-surface properties, rainfall characteristics, and evaporative demand – conditions that humans are modifying at an unprecedented scale (e.g., Crutzen, 2002; Dore, 2005; Gordon et al., 2005; Rockström et al., 2009b; Trenberth, 2011). Understanding evaporation interaction with land and climate is essential, because evaporation holds a key role in regulating hydrological flows as well as atmospheric feedback. One important land-atmosphere mechanism is the contribution of terrestrial evaporation to precipitation through the process of moisture recycling, which has implications for both water and land management. For example, studies have shown that changes in land-use may potentially reduce crop yields through reductions in moisture recycling (Bagley et al., 2012), that irrigation may increase moisture recycling (e.g., Tuinenburg, 2013; Wei et al., 2013), and that livelihoods in some semi-arid regions are particularly vulnerable to changes in upwind moisture source regions (Keys et al., 2012).

Up to now, moisture recycling studies have only analysed total evaporation. However, the partitioning between transpiration, vegetation interception, floor interception, soil moisture evaporation, and open water evaporation depend on land-use and meteorological conditions. For example, interception and soil moisture evaporation are ephemeral (Gerrits et al., 2009), whereas transpiration continues long into the dry season depending on infiltration rates and the capacity of the soil in the root zone to retain moisture. Vegetation that can access deeper soil moisture can therefore maintain evaporation through transpiration beyond what can be sustained
by interception alone. Another example is that transpiration ratios (i.e., transpiration as part of total evaporation) can be relatively higher in wet years (compared to dry years), but smaller in wet months (compared to dry months) (Savenije 2004). The reason is that wet months tend to have high interception preceding transpiration and consuming the already limited energy available for evaporation, whereas wet years tend to receive increased rainfall during the rainy season that stores and transpires into the dry season. Savenije (2004) suggested that these temporal differences of different evaporation fluxes would have different moisture recycling patterns.

Earlier studies of evaporation time scales have analysed the role of soil moisture for drought (e.g., Serafini and Sud 1987; Delworth and Manabe 1988), the precipitation persistence in climate modelling (e.g., Koster and Suarez 1996), as well as the evaporation response time scale to drying soils (e.g., Teuling et al. 2006) and for inter-comparing and improving land surface models (e.g., Lohmann and Wood 2003; Wang et al. 2006). Scott et al. (1997) described the timescale of evaporation response through convolution representation of precipitation history and applied it on interception, soil evaporation and transpiration globally. Lohmann and Wood (2003) employed a similar approach to compare 16 land surface models and found significant differences in response between models. Nevertheless, the role of evaporation partitioning and evaporation time scales specifically for moisture recycling has not been studied.

Although there have been much efforts in estimating global land evaporation and evaporation partitioning, the actual magnitudes of the different evaporative fluxes remain disputed. Methods to estimate spatially distributed global land evaporation can broadly be grouped into land surface models, remote sensing, reanalysis, and data-upscaling methods. While the latter two generally do not provide evaporation partitioning, the first two methods are highly reliant on the assumed parameters, algorithms, and terminology definitions in order to assess the partitioning. Thus, it is not surprising that the range of reported evaporation partitioning is large. Model-based global mean transpiration ratio estimates range from 38 to 80 % (see Sect. 5 and Table 3).

Validation of spatially and temporally distributed global evaporation partitioning data is challenging, as observational measurements are constrained in space and time, and suffer from uncertainties themselves. Although eddy covariance measurements have often been used in validating modelled total evaporation (e.g., Liu et al. 2012; Miralles et al. 2013; van den Hoof et al. 2013; Bagley et al. 2011) and sporadically used for deriving evaporation (e.g., Jung et al. 2010) and evaporation partitioning (e.g., Czikowsky and Fitzjarrald 2009), there are still many issues to be resolved: e.g., non-closure of energy balance, location bias, and upscaling (e.g., Twine et al. 2000; Wilson et al. 2002; Chen et al. 2011; Xiao et al. 2012). A combination of isotope measurement techniques and satellite observations were recently used to investigate evaporative partitioning at the river basin and global scale (Jasechko et al. 2013, 2014), leading to high and disputed (Coenders-Gerrits et al. 2014; Schlaepfer et al. 2014; Sutanto et al. 2014) estimates of the transpiration ratio (80–90 %) (see also Sect. 5). In addition, research initiatives such as GEWEX LandFlux-EVAL and ESA WACMOS-ET (e.g., Jiménez et al. 2011; Miralles et al. 2013) that accumulate knowledge through inter-comparing evaporation and evaporation partitioning are still ongoing.

Thus, there remain many difficulties and uncertainties in estimating evaporation partitioning. In particular, the lack of evaporation partitioning data available at the spatial and temporal scale required for moisture tracking might be a reason for the omission of moisture recycling research in the potentially contrasting effects of separated evaporation fluxes.

The research presented here is divided into two separate research papers. The general aim is to investigate the characteristics and roles of different evaporation fluxes to the atmosphere with respect to moisture recycling. This paper (Part 1) analyses the temporal characteristics of partitioned evaporation on land, and presents and evaluates STEAM (Simple Terrestrial Evaporation to Atmosphere Model) — a hydrological land surface model developed and used for the analyses, van der Ent et al. (2014). (hereafter, Part 2), tracks interception and transpiration fluxes in the atmosphere using the WAM-2layers (Water Accounting Model 2-layers) and investigates the resulting moisture recycling patterns.

Specific research questions investigated in this paper relate to the temporal characteristics important for understanding the reasons for evaporation fluxes to produce different moisture recycling patterns: 1) what are the terrestrial residence time scales of evaporation fluxes? 2) how does the timing of precipitation matter for evaporation partitioning? 3) how robust are the temporal characteristics to uncertainties in storage capacities? We use STEAM to model these fluxes. As a relatively simple evaporation model for analysing the relationship between land-use and moisture recycling, STEAM aims to 1) be tailored for coupling with the atmospheric moisture recycling model WAM-2layers, 2) be flexible for land-use change by land-use parametrisation and by including representation of features particularly important for evaporation (e.g., phenology and irrigation), 3) remain simple, transparent, and computationally efficient, and 4) simulate evaporation and evaporation partitioning in line with current knowledge.

2 Model description

STEAM (Simple Terrestrial Evaporation to Atmosphere Model) is a process-based model assuming water balance at grid cell level. Because of our need to properly quantify partitioned evaporation and its seasonal variations, STEAM includes an irrigation module and calculates dynamic seasonal vegetation parameters based on meteorological condi-
tions. For our current research purposes, we have considered it acceptable to disregard groundwater interactions and lateral flows.

STEAM estimates five evaporative fluxes, and is represented by five stocks, see Fig. [1]. First, the vegetation interception stock $S_v$ represents canopy and vegetation surface evaporation. Water intercepted by the canopy and vegetation surface $S_v$ is vegetation interception $E_v$, and the water exceeding the storage capacity $S_{v,max}$ is throughfall $P_{tf}$. Second, the floor interception stock $S_f$ represents the ground and litter surface which intercepts the throughfall. The interception from this stock is floor interception $E_f$. The remainder is effective precipitation $P_{eff}$, which is generated when the storage $S_{f,max}$ is exceeded. Third, water that subsequently reaches the unsaturated root zone stock $S_{uz}$ can be evaporated either as soil moisture evaporation $E_{sm}$ or be taken up by plant roots and transpire as transpiration $E_t$. Fourth, the water stock $S_w$ represents open water in the land-use classes water (01:WAT) and wetlands (12:WET), and water below vegetation in the land-use classes wetlands (12:WET) and rice paddies (19:RIC). The water stock is replenished by adding water $J_{add}$ that prevents dry-out in the absence of lateral flow routines. Water below vegetation also receives $P_{tf}$ from vegetation. Excess water comprises $Q_{uz}$ (exceeding $S_{uz,max}$) from the unsaturated zone and $Q_{w}$ from the water stock (exceeding $S_{w,max}$). The last and fifth stock $S_{snow}$ does not have a limit, and allows snowfall $P_{tf}$ to accumulate until melting occurs. Snowmelt $P_{melt}$ is allowed only if there is snow in $S_{snow}$. If the daily mean temperature $T_{mean}$ is above 273 K, $P_{melt}$ goes directly to the floor interception stock, otherwise it only adds to $Q_{uz}$. In case of irrigation, some water is assumed to be spilled to the vegetation $I_v$, the floor $I_f$ and the water bodies $I_w$. All notations are listed in Appendix A.

### 2.1 Potential evaporation

Total evaporation, the sum of vegetation interception $E_v$, floor interception $E_f$, transpiration $E_t$, soil moisture evaporation $E_{sm}$, and open water evaporation $E_{wp}$, is driven by the daily potential evaporation, and restricted by resistances and water availability. The Penman–Monteith equation (Monteith 1965) is used to estimate the daily potential evaporation $E_{p,day}$ [m m$^{-2}$ d$^{-1}$], which is formulated as follows:

$$E_{p,day} = \frac{\delta(R_{net} - G) + \rho_a C_p D_s/\rho_a}{\rho_w \lambda (\delta + \gamma)} \tag{1}$$

where $\delta$ [kPa K$^{-1}$] is the gradient of the saturated vapour pressure function, $R_{net}$ [MJ m$^{-2}$ d$^{-1}$] is the net radiation, $G$ [MJ m$^{-2}$ d$^{-1}$] is the ground heat flux, $\rho_a$ [kg m$^{-3}$] is the density of air, $C_p$ [1.01 $\times$ 10$^{-3}$ MJ kg$^{-1}$ K$^{-1}$] is the specific heat of moist air at constant pressure, $D_s$ [kPa] is the vapour pressure deficit, $\rho_w$ [kg m$^{-3}$] is the density of water, $\lambda$ [MJ kg$^{-1}$] is the latent heat of water vapourisation, $\gamma$ [kPa K$^{-1}$] is the psychometric constant, and $r_a$ [d m$^{-1}$] is the aerodynamic resistance. Note that $r_a$ is represented by $r_{a,v}$ for vegetation, $r_{a,f}$ for floor and $r_{a,w}$ for water. The calculations of $\delta$, $R_{net}$, $G$, $D_s$, $\lambda$, $\gamma$ and the different $r_a$ are given in Appendix B. The potential evaporation $E_{p,day}$ in Eq. (1) does not include surface stomatal resistance $r_{s,sm}$ for transpiration or surface soil moisture resistance $r_{s,sm}$ for soil moisture evaporation. Thus, we introduce $k$ (used in Eq. 8, 10, and 11), which is expressed as a function of a surface resistance $r_s$ and an aerodynamic resistance $r_a$:

$$k(\frac{r_s}{r_a}) = \left(1 + \frac{\gamma}{r_s + \gamma}\right)^{-1} \tag{2}$$

The surface stomatal resistance $r_{s,sm}$ is calculated based on the Jarvis–Stewart stress function and optimal temperature based on latitude and altitude, see Appendix B for details. The soil moisture resistance $r_{s,sm}$ is applied to soil moisture evaporation and estimated based on the soil moisture content of the top soil layer (Bastiaanssen et al. 2012).

$$r_{s,sm} = r_{s,sm,\min} \Theta_{top}^{-3} \tag{3}$$

where $r_{s,sm,\min}$ is the minimum surface soil moisture resistance assumed as $3.5 \times 10^{-4}$ d m$^{-1}$, and $\Theta_{top}$ [-] is the effective saturation expressed as:

$$\Theta_{top} \equiv \frac{\Theta_{top, n} - \Theta_{top, res}}{\Theta_{top, sat} - \Theta_{top, res}} \tag{4}$$

Since there is no explicit top soil storage in STEAM, top soil moisture at the present time $\Theta_{top, n}$ [-] is derived daily, based on the inflow to the unsaturated storage and top soil moisture from the previous day $\Theta_{top, n-1}$ (Pellarin et al. 2013):

$$\Theta_{top, n} = \Theta_{top, n-1} e^{-\Delta n/\chi} + (\Theta_{sat} - \Theta_{top, n-1}) (1 - e^{-\rho_{top}/\gamma}) + \Theta_{top, res} \tag{5}$$

where $\Delta n$ is the time step of 24 h, $\Theta_{top, res}$ is the volumetric residual soil moisture content assumed as 0.01, $\rho_{top}$ is the top soil depth, and $\chi$ is the dry out parameter which varies with clay content of the top soil. The assumed $\rho_{top}$ is 0.03 m. In Pellarin et al. (2013), the values used for $\rho_{top}$ were 0.05 m and 0.1 m, but we considered that a shallower depth is more relevant for estimating soil moisture evaporation stress. The dry out parameter $\chi$ is estimated using the following semi-empirical equation:

$$\chi = \frac{\rho_{top}}{0.1} \max[\chi_{min}, 32 \ln(\eta_{clay} + 174)] \tag{6}$$

where $\eta_{clay}$ is the clay content [%] and $\chi_{min}$ is the minimum of $\chi$ taken as 60 h. This set of equations (Eq. 5) and 6 was tested in semi-arid West Africa, in the type of regions where soil moisture evaporation is most important. Factors not taken into account include solar radiation, the presence of vegetation and the wind velocity (Pellarin et al. 2013).
2.2 Actual evaporation

To simulate actual evaporation at 3 hour time steps ($\Delta t$), we first downscale the daily potential evaporation $E_{p,\text{day}}$ using the diurnal distribution of ERA-I 3 h evaporation. The down-scaled potential evaporation is subsequently used to evaporate moisture in the following logical sequence — vegetation interception, transpiration, floor interception, and soil moisture evaporation:

$$E_{v, \text{lu}, vs} = E_{v, \text{lu}, vw} = \min \left( \frac{S_{v, \text{lu}}}{\Delta t}, E_p (r_{a, v}) \right) \tag{7}$$

$$E_{t, \text{lu}, vs} = \min \left( \frac{S_{a, \text{lu}}}{\Delta t}, \max \{0, [E_p (r_{a, v}) - E_{v, \text{lu}, vs}] \cdot k (r_{a, v}, r_{s, \text{sat}}) \} \right) \tag{8}$$

$$E_{t, \text{lu}, vs} = \min \left( \frac{S_{a, \text{lu}}}{\Delta t}, \max \{0, E_p (r_{a, v}) - E_{v, \text{lu}, vs} - E_{t, \text{lu}, vs} \} \right) \tag{9}$$

$$E_{w, \text{lu}, vs} = \min \left( \frac{S_{w, \text{lu}}}{\Delta t}, r_{s, \text{sat}} \right) \tag{10}$$

where the first subscript ($v$, $t$, or $w$) denotes an individual evaporative flux, the second subscript ($\text{lu}$) the land-use type ID (see Table C1), and the third subscript ($\text{vs}$, $vw$ or $ow$) the type of vegetation-water occupancy (see Table C2).

$$\alpha = \max \{0, [E_p (r_{a, t}) - E_{v, \text{lu}, vs} - E_{t, \text{lu}, vs} - E_{w, \text{lu}, vs}] \cdot k (r_{a, t}, r_{s, \text{sat}}) \} \tag{11}$$

$$E_{w, \text{lu}, vw} = \min \left( \frac{S_{w, \text{lu}}}{\Delta t}, \max \{0, E_p (r_{a, w}) - E_{v, \text{lu}, vw} - E_{t, \text{lu}, vw} \} \right) \tag{12}$$

For the water land-use type and the fraction of open water $\phi_{ow}$ in wetlands, evaporation is expressed as:

$$E_{w, \text{lu}, ow} = \min \left( \frac{S_{w, \text{lu}}}{\Delta t}, \max \{0, E_p (r_{a, w}) \} \right) \tag{13}$$

The total of an evaporation flux from wetland (12:WET) or rice paddy (19:RIC) is determined by the weighted sum based on the fractions of vegetation covered soil $\phi_{vs}$, vegetation covered water $\phi_{vw}$, and open water $\phi_{ow}$ (see also Table C2):

$$E_{j, \text{lu}} = \phi_{lu} \cdot E_{v, \text{lu}, vs} + \phi_{lu} \cdot E_{v, \text{lu}, vw} + \phi_{lu} \cdot E_{w, \text{lu}, ow} \tag{14}$$

where $E_{j, \text{lu}}$ is an evaporation flux ($j$ denotes $v$, $t$, $f$, $sm$, or $w$) of the land-use type $\text{lu}$.

Subsequently, the total of an evaporation flux from a grid cell is determined by the weighted sum of the land-use types:

$$E_{j} = \sum_{\text{lu}=1}^{\text{lu}=19} \phi_{lu} E_{j, \text{lu}} \tag{15}$$

where $\phi_{lu}$ is the land-use occupancy fraction of the land-use type $\text{lu}$.

2.3 Phenology

The growing season index $i_{GS}$ (Jolly et al., 2005) varies between 0 and 1, and is used to determine the seasonal variations of leaf area $i_{LA}$. We formulate $i_{GS}$ in STEAM as follows:

$$i_{GS} = f \left( T_{\text{min}} \right) \cdot f \left( N \right) \cdot f \left( \theta_{uz} \right) \tag{16}$$

where $f(T_{\text{min}})$ is the stress function of minimum temperature, $f(N)$ is the stress function of day length, and $f(\theta_{uz})$ is the stress function of soil moisture. Note that $f(\theta_{uz})$ is a modification of the original expression for $i_{GS}$, where vapour pressure deficit $D_a$ was used as a proxy for soil moisture (Jolly et al., 2005). However, since soil moisture is calculated in STEAM, it makes sense to use the soil moisture stress function to replace the original vapour pressure stress function.

The stress functions are expressed as:

$$f \left( T_{\text{min}} \right) = \begin{cases} 0 & T_{\text{min}} \leq T_{\text{min}, \text{low}} \\ \frac{T_{\text{min}} - T_{\text{min}, \text{low}}}{T_{\text{min}, \text{high}} - T_{\text{min}, \text{low}}} & T_{\text{min}, \text{high}} > T_{\text{min}} > T_{\text{min}, \text{low}} \\ 1 & T_{\text{min}} \geq T_{\text{min}, \text{high}} \end{cases} \tag{17}$$

$$f \left( \theta_{uz} \right) = \begin{cases} 0 & \theta_{uz} \leq \theta_{uz, \text{wp}} \\ \frac{\theta_{uz} - \theta_{uz, \text{wp}}}{\theta_{uz, \text{wp}} - \theta_{uz, \text{wp} + c_{\text{uz}}}} & \theta_{uz, \text{wp}} < \theta_{uz} < \theta_{uz, \text{fc}} \\ 1 & \theta_{uz} \geq \theta_{uz, \text{fc}} \end{cases} \tag{19}$$

where the lower sub-optimal minimum temperature $T_{\text{min}, \text{low}}$ is $271.15 \, \text{K}$, and the higher $T_{\text{min}, \text{high}}$ is $278.15 \, \text{K}$. The lower sub-optimal threshold day length $N_{\text{low}}$ is assumed to be $36,000 \, \text{s}$, and the higher $N_{\text{high}}$ is $39,600 \, \text{s}$ (Jolly et al., 2005). $T_{\text{min}}$ is taken from the coldest 3 h ERA-I temperature of the day. Calculation of day length $N$ follows the approach of Glarner (2006). The soil moisture stress parameter $c_{\text{uz}}$ is fixed at 0.07 (Matsumoto et al., 2008). The soil moisture content $\theta_{uz}$ is $S_{uz}/y_{uz}$, where $y_{uz}$ [m] is the depth of the unsaturated root zone. The soil moisture contents at wilting point $\theta_{uz, \text{wp}}$ and at field capacity $\theta_{uz, \text{fc}}$ depend on soil type.
vegetation stock linked to vegetation types due to the inherent differences in vegetation vary with vegetation type and a strong relationship has not yet been established. In fact, Breuer et al. (2003) even suggests that no general relationship can be established across km and grow to almost 28 000 where c varies with vegetation type and a strong relationship has not

2.4 Storage capacities

The storage capacity determines the maximum water availability for the evaporation flux of concern. We derived vegetation interception storage capacity \( S_{i LS, max} \) from the monthly \( i_{LA} \) based on the storage capacity factor \( c_{sc} \) of roughly 0.2 reported by, for example, de Jong and Jetten (2007) and used in van den Hoof et al. (2013):

\[
S_{v, max} = c_{sc} \cdot CAR \cdot i_{LA}, \tag{21}
\]

where \( CAR \) is the area reduction factor introduced to compensate for rainfall heterogeneity in space and time. The relationship between \( i_{LA} \) and vegetation interception storage varies with vegetation type and a strong relationship has not yet been established. In fact, Breuer et al. (2003) even suggests that no general relationship can be established across vegetation types due to the inherent differences in vegetation structures. Nevertheless, vegetation stock linked to \( i_{LA} \) has proven to be useful in many cases where there is a lack of detailed vegetation information.

We assume \( CAR \) to be 0.4 for STEAM running with a 3 h time step at the 1.5’ scale. Area reduction factors have been developed to establish a relationship between average precipitation and extreme precipitation of a region, but can be analogously used to reduce interception storage capacity. In an example diagram obtained from catchment analyses (Shuttsworth, 2012), areas larger than 10 000 km² have an area reduction factor up to approximately 0.6. In STEAM, grid cell areas with 1.5’ resolution are 10 000 km² already at 68° N and grow to almost 28 000 km² at the equator. Ideally, \( CAR \) should vary with the area considered and rainfall duration, but due to a lack of well-established functions, we consider \( CAR = 0.4 \) to be acceptable.

The floor interception storage capacity \( S_{f, max} \) [m] is modelled as a function of the leaf area and a certain base value:

\[
S_{f, max} = c_{sc} \cdot CAR \cdot [1 + 0.5 \cdot (i_{LA, max} + i_{LA, min})]. \tag{22}
\]

The floor storage capacity increases in areas with vegetation, due to litter formation from fallen leaves. A base value is considered, because wetting of the surface always occurs irrespective of the land cover. However, litter is assumed to have been removed in croplands (i.e., 13:CRP, 15:MOS, 18:IRR, and 19:RIC). Thus, \( S_{f, max} \) [m] for crops corresponds to that of the litter-free floor:

\[
S_{f, max, crops} = c_{sc} \cdot CAR. \tag{23}
\]

As a result of the large grid scale (reflected in the area reduction factor), interception storage in STEAM is smaller than normally found in point scale field studies. For example, the vegetation interception storage capacity at the maximum \( i_{LA} \) of 5.5 is 0.44 mm, which is about a third of the 1.2 mm reported in a summer temperate forest (Gerrits et al., 2010) and a fraction of the 2.2 to 8.3 mm per unit of crown projected area in a tropical rainforest site (Herwitz, 1985).

The storage capacity of the unsaturated root zone \( S_{uz, max} \) is assumed to reach field capacity when:

\[
S_{uz, max} = \theta_{fc} \cdot y_{uz}. \tag{24}
\]

The \( S_{uz, max} \) is modelled as a function of soil texture and land-use based rooting depth. This is a simplification as many other factors govern root water uptake, including topography (Gao et al., 2013), soil properties, hydraulic redistribution of soil water by roots (Lee et al., 2005), groundwater table (Miguez-Macho and Fan, 2012), and climate (Peddes et al., 2001). In addition, variations of rooting distribution (e.g., Jackson et al., 1996) and the existence of deep roots (e.g., Canadell et al., 1996; Kleidon and Heimann, 2000) may conflict with the assumption of one rooting depth parameter per land-use type.

2.5 Irrigation

STEAM includes irrigation because it has been shown to constitute an important moisture source to the atmosphere (e.g., Gordon et al., 2005; Lo and Famiglietti, 2013; Tuinenburg, 2013; Wet et al., 2013). Irrigation water supplied is assumed to meet the irrigation requirement and is not restricted by water availability. Net irrigation enters the unsaturated zone and is estimated as a function of soil moisture. In rice paddies (19:RIC), irrigation water simply upholds a 10 cm water level. For non-rice crops (18:IRR), irrigation requirement \( I_{req} \) is the amount of water needed to reach field capacity in the unsaturated root zone:

\[
I_{req} = \max \left[ 0, \frac{y_{uz} (\theta_{uz, fc} - \theta_{uz}) - S_{uz, lu}}{\Delta t} \right]. \tag{25}
\]

However, because a certain amount of irrigation water applied is always lost due to inefficiencies in the system, an irrigation efficiency should be applied in order to correctly estimate runoff and water withdrawal. In STEAM, we assume the gross irrigation \( I_g \) to be twice the \( I_{req} \). Although irrigation efficiency in practice varies greatly with irrigation technique, crop type and country (Rohwer et al., 2007), we consider our simplification acceptable since the gross irrigation assumption affects evaporation (our major concern) less than, e.g., runoff and water withdrawal. Of gross irrigation applied to irrigated non-rice crops (18:IRR), 15 % is directed to the vegetation interception stock \( S_v \), and 85 % to the floor interception stock \( S_f \). Of the gross irrigation applied to rice paddies (19:RIC), 5 % is directed to vegetation interception stock \( S_v \), 5 % to the floor interception stock \( S_f \) (assuming inter-paddy pathways), and 90 % to the water stock \( S_w \).
### 3 Data

Meteorological data were taken from the ERA-Interim reanalysis (ERA-I) produced by the European Centre for Medium-Range Weather Forecasts (ECMWF) \(^{[1]}\). We used evaporation, precipitation, snowfall, snowmelt, temperature at 2 m height, dew point temperature at 2 m height, wind speed in two directions at 10 m height, incoming shortwave radiation, and net longwave radiation. All meteorological forcings are given at 3 h and 1.5° latitude x 1.5° longitude resolution. The data used covers latitudes from 57° S to 79.5° N for the years 1985–2009.

The monthly varying land-surface map used in STEAM consists of 19 land-use types, see Table \(\text{C1}\). The first 17 International Geosphere Biosphere Programme (IGBP) land-use types are based on the Land Cover Type Climate Modeling Grid (CMG) MCD12C1 created from Terra and Aqua Moderate Resolution Imaging Spectroradiometer (MODIS) data \(^{[2]}\) for the year 2001. The two irrigated land-use classes are based on the dataset of Monthly Irrigated and Rainfed Crop Areas around the year 2000 \(^{[3]}\) (MIRCA2000) V1.1. \(^{[4]}\). The resolution for MODIS is 0.05° and for MIRCA2000 0.5’. To create the joint map, monthly irrigated land from MIRCA2000 were taken to replace primarily MODIS cropland fraction (13:CRP), and secondarily MODIS cropland/natural mosaic fraction (15:MOS). The joint map has a total land area of 133,146,465 km² and includes inland waters except big lakes.

Soil texture data has been taken from the Harmonized World Soil Database (HWSD) \(^{[5]}\) (FAO/IASA/ISRIC/ISSCAS/JRC 2012) and we assigned volumetric soil moisture content at saturation, field capacity and wilting point based on the United States Department of Agriculture (USDA) soil classification \(^{[6]}\) (Saxton and Rawls 2006). Top soil saturation, subsoil field capacity and wilting point have been assigned to the original 30” resolution, and scaled up to 1.5° by area weighing.

For evaporation evaluation, we used the Landflux-EVAL evaporation benchmark products \(^{[7]}\) (Mueller et al. 2013) for the years 1989–2005. This data product consists of a merged synthesis from 5 satellite or observation-based datasets, 5 land-surface model simulations, and 4 reanalysis datasets. For runoff evaluation, composite and model runoff fields from the Global Runoff Data Centre (GRDC) were used \(^{[8]}\) (Fekete et al. 2000). The model runoff fields are the simulations of the GRDC Water Balance Model (GRDC-WBM), whereas the composite runoff fields (GRDC-Comp) are the model runoff corrected by observed station discharge \(^{[9]}\) (Fekete et al. 2000). In addition, we also used ERA-I runoff fields \(^{[10]}\) (Balsamo et al. 2011) in our comparison. It should be noted that the ERA-I runoff fields form a separate dataset that does not directly correspond to ERA-I precipitation minus evaporation. The river basin map is based on the global 30-min drainage direction map of Doll and Lehner \(^{[11]}\) (2002).

### 4 Methods

#### 4.1 Model evaluation

The model evaluation comprises the following model output: total and land-use based evaporation, total and land-use based evaporation partitioning, runoff, irrigation, and irrigation evaporation contribution. Total global fluxes are calculated based on a land area of 133,146,465 km² (including Greenland and excluding Antarctica) and for the years 1999–2008. Land-use evaporation is obtained from Eq. \(14\). Irrigation evaporation contribution was calculated based on the difference in evaporation between STEAM simulations with and without the irrigation routine turned on. Runoff \(Q\) from STEAM has been derived from subtracting mean evaporation and mean snow storage changes from mean precipitation:

\[
Q = P - E - \frac{dS_{\text{snow}}}{dt}.
\]

Snow storage changes were subtracted because snow accumulated in glaciers may carry over storage from year to year. Otherwise, most storage changes may be neglected at an annual time scale. Then runoff comparison includes two additional STEAM scenarios: one simulation without irrigation (because irrigation is not always included in land surface models), and one with 5% uniform reduction in precipitation forcing (because ERA-I precipitation forcing is higher than several other precipitation datasets, see Appendix \(\text{D}\).

#### 4.2 Characterisation of partitioned evaporation fluxes

##### 4.2.1 Time scales of evaporation fluxes

The time scales \(\tau_{\text{ts}}\) of the evaporation fluxes is defined as the mean stock over the mean flux rate of concern \(j\):

\[
\tau_{\text{ts}, j} = \frac{S_j}{E_j}.
\]

Figure \(1\) shows the stock of origin for each evaporation flux. Because both \(E_{\text{sm}}\) and \(E_{\text{i}}\) come from \(S_{\text{az}}\), we assumed a stock of soil moisture evaporation \(S_{\text{az,sm}}\) and a stock of transpiration \(S_{\text{az,1}}\). To obtain \(S_{\text{az,sm}}\), we multiplied \(\theta_{\text{top}}\) with the assumed top soil depth \(y_{\text{top}}\). To obtain the stock \(S_{\text{az,1}}\), \(S_{\text{az,sm}}\) was subtracted from the total water available in the unsaturated zone \(S_{\text{az}}\):

\[
S_{\text{az,1}} = S_{\text{az}} - S_{\text{az,sm}} = \theta_{\text{az}} y_{\text{az}} - \theta_{\text{top}} y_{\text{top}}.
\]

Because the time scale becomes infinite when the flux approaches zero, time scales are not given for areas where the mean evaporation flux is below 0.01 mm d⁻¹. Coastal areas where the land area fraction is less than 100% were removed from the time scale analysis. The time scale for open water evaporation was not calculated.
4.2.2 Evaporation partitioning: time since precipitation

We are interested in how evaporation partitioning evolves with time after precipitation ceases. To do this, we grouped each grid cell at every time step in categories depending on the time that has past since precipitation. Grid cells at a certain time step that has not received precipitation since n time steps back are placed in the (n+1)th category. Subsequently, evaporation partitioning for each category was retrieved from the model simulation.

In addition, the importance of the evaporation partitioning in relation to rainfall also depends on the evaporated quantity. Thus, we present the portion of evaporation flux during rainy or dry conditions by using evaporation efficiencies \( \beta_{\text{wet}} \) and \( \beta_{\text{dry}} \) as measures:

\[
\beta_{\text{wet},j} = \frac{\sum E_{\text{wet},j}}{\sum E_j}, \quad \beta_{\text{dry},j} = \frac{\sum E_{\text{dry},j}}{\sum E_j}.
\]

Here, \( \beta_{\text{wet}} \) represents the mean annual portion of an evaporation flux that evaporates during a 3 hour time step with precipitation, and \( \beta_{\text{dry}} \) represents the mean annual portion of an evaporation flux that evaporates after experiencing more than 24 hours of no precipitation. To qualify as a wet time step, a 3 hour time step must have \( >0.01 \text{ mm precipitation} \).

The subscript \( j \) denotes the evaporation flux of concern. Construction of these evaporation efficiency measures is useful for answering questions such as: how much of total vegetation interception occurs during rainy periods?

4.2.3 Robustness

Large uncertainties exist in evaporation partitioning and estimation of storage capacities. To verify how robust or sensitive the temporal characteristics are to these uncertainties, we performed a sensitivity analysis with two scenarios: transpiration-plus and interception-plus. In transpiration-plus, the unsaturated zone storage capacity increased by 20 % and the vegetation and floor interception storage capacity decreased by 50 %. In interception-plus, the increase and decrease in the storages are reversed, see Table 5.

5 Results: model evaluation

5.1 Total evaporation

STEAM estimates global annual terrestrial evaporation as 555 mm-year\(^{-1} \) (i.e., 73,900 km\(^3\)year\(^{-1}\), spatial distribution is shown in Fig. 3. This is comparable to current global evaporation datasets. In the Water Model Intercomparison Project (WaterMIP), the range of evaporation given by eleven models was 415–585 mm-year\(^{-1} \) for the period 1985–1999. By subtracting global runoff from precipitation products for the years 1984–2007, Vinukollu et al. (2011) arrived at global evaporation rates of 488–558 mm-year\(^{-1} \).

In the LandFlux-EVAL multi-data set synthesis, the global mean evaporation was 493 mm-year\(^{-1} \) as given by a combination of land-surface model simulations, observational dataset, and reanalysis data for both the period of 1989–1995 and 1989–2005 (Mueller et al. 2013).

Figure 3 shows how STEAM evaporation compares to the LandFlux-EVAL product for 1989–2005. STEAM evaporation is within the inter-quartile range of all LandFlux-EVAL products in the tropics, the United States, parts of Europe, South Asia, northern Russia and large parts of Africa south of Sahel. The upper quartile is mostly exceeded in the boreal forests in the northern latitudes, China, Argentina and the Sahel. Most exceedance of STEAM evaporation is in comparison with the land surface models, and the least with the reanalyses data included in the LandFlux-EVAL product. Only a few limited patches in northern Canada, Sudan, Argentina and northern China exceed the LandFlux-EVAL maximum. Seasonally, Fig. 3 shows that Northern Hemisphere spring and summer are generally more in range compared to winter and fall, when STEAM tends to have higher evaporation rates in the northernmost latitudes compared to LandFlux-EVAL.

However, LandFlux-EVAL excluded some high evaporation values in the northern latitudes based on physical constraints (Mueller et al. 2013), which consequently eliminates potentially important winter time interception (Schläpfer et al. 2014).

Evaporation contributions per land-use type are listed in Table 1 and compared to the other studies in Table 2. The highest evaporation rates are found in irrigated lands, evergreen broadleaf forests, and open waters. This is followed by wetlands, savannas, deciduous broadleaf forests, natural mosaics, woody savannas, mixed forests, and rainfed croplands. Evaporation rates in the lower tier include contributions from needleleaf forests, grasslands, and shrublands. In general, STEAM evaporation is comparable to the estimates of Gordon et al. (2005), the compilation results of Schlesinger and Jasechko (2014) (based on Mu et al. 2011), and the field data from Rockström et al. (1999). The mixed forest evaporation estimate in STEAM is double that of Gordon et al. (2005), but the area is also very different, suggesting substantial differences in forest definition. Closed shrublands in STEAM also produces higher evaporation rates, but because the numbers are for shrublands in general and not closed shrublands in particular, the shrublands comparison is inevitably inconclusive. Some caution is warranted in comparing evaporation rates across studies. Nevertheless, this comparison shows that evaporation estimates in STEAM are within the range of previous estimates.

5.2 Evaporation partitioning

In STEAM, the dominating evaporation flux is transpiration \( E_t \) (59 %), followed by vegetation interception \( E_v \) (21 %), floor interception \( E_f \) (10 %), soil moisture evaporation \( E_{sm} \)
(6 %) and lastly, open water evaporation $E_w$ (4 %). The global distribution of the annual mean evaporation fluxes is shown in Figs. 2 and 5 (as percentage of total evaporation). Seasonal variations of evaporation fluxes are shown over latitudes in Fig. 6. It is shown that transpiration dominates in the densely vegetated areas in the tropics. In addition, transpiration rates increase over the boreal forests during the Northern Hemisphere summer.

Table 3 provides an overview of evaporative partitioning values in the literature and in STEAM. We note that the STEAM global mean transpiration ratio is in good agreement with the literature compilation presented by Coenders-Gerrits et al. (2014) and the LPJ estimate by Gerten et al. (2005), but higher than other land-surface model simulations (Alton et al., 2009; Lawrence et al., 2007; Choudhury et al., 1998; Dirmeyer et al., 2006; Jasechko et al., 2013) estimated the transpiration ratio to be 80–90 % using a combination of isotope measurement techniques and satellite observations at river basin and the global scales. However, their results have been challenged by Coenders-Gerrits et al. (2014) who showed that the transpiration ratio reduces to 35–80 % by using other input data, Schlesinger and Jasechko (2014) who estimated the global transpiration ratio to be 61 % based on literature data compilation, and by Schlaepfer et al. (2014) who argued that Jasechko et al. (2013)'s underlying assumption that isotope ratios of a lake would be representative for the entire catchment is flawed. A number of possible explanations for the high transpiration ratio bias in isotope studies is also offered by Sutanto et al. (2014).

Table 1 shows the annual average evaporation fluxes as a percentage of total evaporation per land-use class. Transpiration is the dominant evaporation flux in almost all land-use types: 50–64 % in forests, 61 % in grasslands, 72 % in croplands, and 58–65 % in shrublands. The exceptions are, logically, barren lands (17 %), snow/ice (16 %) and open waters (01:WAT), forested and savannah areas, but are much lower (down to < 30 %) in the western US, India, southeastern China, and South Africa. Alton et al. (2009) report global mean transpiration ratios of 49–65 % in forests, 32–60 % in grassland, and 44–51 % in shrublands. The order of magnitude is similar to STEAM, but transpiration ratios for shrublands are lower. Schlesinger and Jasechko (2014) compiled satellite-based estimates from Mu et al. (2011) and arrived at 70 % transpiration in tropical forests, 55–67 % in other forests, and 57–62 % in grasslands. Choudhury et al. (1998) used a biophysical process-based model, and estimated transpiration ratio to amount to 56–77 % in three rainforest regions, 63–82 % in three savannah regions, and 37–82 % in seven cropland areas. Transpiration for river basins shown in the isotope study of Jasechko et al. (2013) show transpiration ratios above 70 % in grassland-dominated areas in the western United States, van den Hoof et al. (2013) evaluated model performance against sites in temperate Europe, and reported transpiration rates of 47–78 % at eight forest sites, and 59–79 % at three grassland sites. Overall, STEAM falls well in the range of the reported evaporation partitioning ratios.

STEAM estimates the vegetation interception ratio as 18 % of rainfall in evergreen broadleaf forest, 17 % in deciduous broadleaf forest, and 18–20 % in needleleaf forest. In comparison, Miralles et al. (2010) arrived at higher canopy interception in coniferous (22 %) and deciduous forest (19 %) than in tropical forest (13 %) using satellite data analysis and literature review. This, interception ratios are comparable, except for tropical forest. In an interception scheme comparison study, Wang et al. (2007) found that taking rainfall type into account increased the performance and decreased interception in the tropics in comparison to the default CLM3 (Community Land Model version 3) interception scheme. Although STEAM uses an area reduction factor to scale interception, this may simply not be enough in the tropical, convective rainfall regimes. On the other hand, field studies have shown high interception ratios in the tropics. For example, Cuartas et al. (2007) reported 16.5 % for two years in Central Amazon, Franken et al. (1992) reported 19.8 % in Central Amazon, and Tobon Marin et al. (2000) reported 12–17 % in Colombian Amazon over four years. Interestingly, Cuartas et al. (2007) also showed that the differences in dry and normal years can differ substantially: 13.3 % in a normal year and 22.6 % in a dry year.

Sensitivity of STEAM evaporation partitioning to precipitation is analysed by a 5 % uniform reduction of precipitation, see Appendix D.

### 5.3 Runoff

STEAM estimates the mean annual global runoff as 43 216 km$^3$ year$^{-1}$ (325 mm year$^{-1}$, 37 % of $P$). Based on discharge data and simulated stream flow simulations, Dai and Trenberth (2002) estimated runoff to be 37 288 ± 662 km$^3$ year$^{-1}$ (35 % of $P$) excl. Greenland and Antarc-
tica). Syed et al. (2010) arrived at 36 055 km$^3$ year$^{-1}$ based on the global ocean mass balance, Oki and Kanae (2006) reported 45 500 km$^3$ year$^{-1}$ including groundwater runoff, and the GRDC composite runoff (GRDC-Comp) is about 38 000 km$^3$ year$^{-1}$ (Fekete et al. 2000). Thus, the STEAM runoff estimate appears to be slightly higher than some of the previous estimates, but lies within the uncertainty range. Differences can partly be explained by the terrestrial area considered in the studies, as well as relatively high $P$ applied (see Appendix D).

STEAM runoff was also compared to GRDC-Comp, GRDC-WBM, and ERA-I runoff data in 13 major river basins of the world, see Figs. 7 and 8a. The largest deviations for both STEAM and ERA-I from the GRDC-Comp runoff are found in the Congo and Nile river basins. However, because Congo precipitation and runoff estimates are particularly uncertain in general (Tshimanga 2012), we cannot evaluate our Congo evaporation estimate based on this specific comparison. As for the Nile river basin, STEAM uses a static land-use map that does not include seasonal variations in wetland size or presence of reservoirs. Since the Nile contains the Sudd, one of the largest wetlands in the world with a highly variable size, evaporation simulation is challenging in this region, even in fine resolution models including complex processes (Mohamed 2005, Mohamed et al. 2007). In several of the northern river basins (e.g., the Mississipi, Mackenzie, and Danube), STEAM runoff is low in comparison to GRDC-Comp. There could be multiple reasons for this underestimation: our simplified snow simulation, our uniform parameterisation of land-use classes across climate zones or simply uncertainties in the forcing data. In support of the latter, the largest uncertainties in evaporation inferred from precipitation and runoff data occur mainly in the higher latitudes (Vinukollu et al. 2011).

Table 3 shows that the STEAM evaporation is close to the mean evaporation provided by the WaterMIP (Water Model Intercomparison Project) (Haddeland et al. 2011, Harding et al. 2011), while both the simulated runoff and the used precipitation forcing is substantially lower. In contrast, in the Lena river basin, STEAM runoff is in range while both evaporation and precipitation have a high bias. In the Amazon basin, the default STEAM simulation slightly overestimates runoff, but reducing precipitation forcing by 5% (see the 95%-$P$ run in Fig. 7) brings runoff down to the level in GRDC-Comp. Also the comparison with WaterMIP indicates that high bias in Amazon precipitation translates into high runoff. This effect of precipitation reduction can also be noted in particularly the Brahmaputra–Ganges, Congo, and Nile river basins. This is not surprising, because precipitation uncertainties have been shown to translate almost entirely into uncertainty in runoff in wet regions, but not at all in arid regions (e.g., Fekete et al. 2004). The relative sensitivity of runoff and evaporation fluxes to precipitation is further accounted for in Appendix D.

The standard deviations between the multiyear mean runoffs in GRDC-Comp (which we here consider as the benchmark runoff) and the other runoffs (GRDC-WBM, ERA-I, and STEAM) are shown in Fig. 8b and c. Among the compared datasets, STEAM runoff deviates the most from GRDC-Comp when Congo is included and the least when Congo is excluded. Note also that omitting irrigation in STEAM increases the runoff deviation to GRDC-Comp, and that reducing precipitation decreases this deviation. Thus, the wet bias in ERA-I precipitation probably explains some of the runoff overestimations we notice in STEAM.

### 5.4 Irrigation

The simulated mean gross irrigation is 1970 km$^3$ year$^{-1}$, and the simulated mean increase in evaporation from irrigation is 1134 km$^3$ year$^{-1}$. The irrigation hotspots in especially India, south-eastern China, and the central US coincide well with where evaporation is enhanced by irrigation input. Our estimates are comparable to previous estimates. Gross irrigation was estimated at 2500 km$^3$ year$^{-1}$ by Döll and Siebert (2002), at 2353 km$^3$ year$^{-1}$ by Seckler et al. (1998), and at 1660 km$^3$ year$^{-1}$ by Rost et al. (2008). The latter study did, however, not take into account recharge to the groundwater. Evaporation contribution by irrigation was simulated at 1100 km$^3$ year$^{-1}$ by Döll and Siebert (2002). While higher evaporation contributions have also been reported in the literature, such as 2600 km$^3$ year$^{-1}$ by Gordon et al. (2005), they could possibly be explained by differences in methods and irrigation maps. Given the uncertainties, the modeling results are considered acceptable in terms of total amounts.

### 6 Results: temporal characteristics

#### 6.1 Terrestrial time scales

The modelled global average time scale (Eq. 27) is 1.3 h for vegetation interception and 7.7 h for floor interception, but 42 days for soil moisture evaporation and 274 days for transpiration in areas with mean evaporation rates higher than 0.01 mm d$^{-1}$. Evaporation rates from vegetation cover and floor are large compared to their respective stocks, resulting in small time scales for interception. In contrast, the stocks in the unsaturated zone are many times larger than the interception stocks, and cause the time scales of soil moisture evaporation and transpiration to extend from days and months. The use of an area reduction factor (see Eq. 21 and 22) leads to interception storage capacities that are smaller in the model than in reality, thus, presumably causing some underestimation of the interception time scales. Nevertheless, the robustness test (Table 5) shows that the magnitude of all evaporation time scales (except for transpiration) are relatively robust against uncertainties in storage capacities.

Figure 9 shows the spatial distribution of mean terrestrial residence time scales (i.e., stock divided by flux) of the par-
tiated evaporation fluxes (Eq. 27). We see that time scales are in general prolonged over the tropics, and over the cold northern latitudes. This finding is consistent with the transpiration response time scale provided by Scott et al. (1997).

Over the tropics, evaporation rates are high, but the stocks are also relatively larger. The time scales of floor and soil moisture evaporation are extended in the tropics, because these fluxes there are suppressed by the relatively high vegetation interception and transpiration.

The temporal variation of the evaporation fluxes at different latitudes is displayed in Fig. 10. Seasonality is distinct for all evaporation fluxes, in particular for transpiration time scales. While the mean latitude transpiration time scale can extend to over 500 days in the mid-latitude winter, it falls well below 100 days in the summer.

Regions and seasons with extremely high transpiration time scales (>300 days) largely coincide with low transpiration in the north, whereas high transpiration rates coincide with intermediate or low time scales (<100 days). On the contrary, relatively high vegetation interception time scales seem positively correlated with high vegetation interception in the tropics, (compare Fig. 2 and 3). This difference can be explained by the limiting factor to evaporation. Transpiration time scales approach infinity as the stock is still wet, whereas vegetation interception time scale often approaches zero when vegetation interception is caused by depletion in vegetation interception stock rather than in evaporative demand. Thus, the high transpiration time scales in the north should be understood as the result of declining evaporative demand, whereas the high vegetation interception time scales in the tropics can be interpreted as the result of a steady and ample supply of precipitation to the vegetation interception stock.

The higher the interception ratios, the lower the evaporation time scales on land (also in consistency with e.g., Scott et al. (1995)), and the faster the overall feedback to the atmosphere. The regions that have a high vegetation interception ratio (Fig 5) coincide with the regions with low atmospheric moisture recycling length scales (van der Ent and Savenije 2011). This suggests that tropical interception is very important for vegetation to maintain atmospheric moisture in the air, and could constitute a large portion of local recycling due to immediate feedback. However, moisture supplied to continents in general (van der Ent et al. 2010), the world’s most important croplands (Bagley et al. 2012), or for rainfall dependent regions (Keys et al. 2012) also relies on remote evaporation sources, which could account for a large part of transpiration. For such cases, upwind modifications that result in changed transpiration rates (e.g., changes in vegetation species, rainwater harvesting practice, CO2 concentrations) may play a larger role for downwind regions than changes in interception. A detailed investigation of the role of interception and transpiration for local and remote moisture recycling is performed in Part 2.

6.2 Evaporation partitioning in relation to time since precipitation

Figure 11 shows the mean latitudinal evaporation ratios by time since precipitation last occurred. Mean latitudinal transpiration ratio is up to 40% during the wet time steps with precipitation, but can amount to up to 90% after just a few dry 3 hour time steps. The largest increase in transpiration ratios with time since precipitation are seen in the cold northern latitudes, where moisture availability is expected to exceed evaporative demand. On the contrary, the vegetation interception ratio is high (up to approximately 60%) during wet time steps, but falls drastically to almost no interception within 6 hours. Similarly to transpiration, soil moisture evaporation ratios generally increase with precipitation-free hours. However, the steepest increase in soil moisture evaporation ratios are found in the equatorial band where the total soil moisture evaporation is very low.

Table 5 shows that transpiration and soil moisture evaporation occur both during wet and dry conditions, whereas vegetation and floor interception evaporation occur almost exclusively during time steps with precipitation. The table shows that 31% of all transpiration occurs during time steps that have endured more than one day of no precipitation, whereas vegetation interception occurs. Instead, 96% of the vegetation interception occurs in a 3 hour time step with precipitation, whereas only 45% of transpiration evaporates in such conditions. Noteworthy is also that these evaporation efficiency numbers (Eq. 29) are robust to changes in the evaporation partitioning: for example, the 96% vegetation interception efficiency persists even when the vegetation interception ratio varies between 12 and 27%. In other words, even with large differences in the evaporation ratio, interception is likely to occur almost exclusively within the wet period, whereas transpiration may have a substantial time lag between the moment water enters the soil and exits through a plant’s stomata. In for example the field study of Farah et al. (2004), transpiration at a tropical woodland site continued for two months after rainfall. This also explains why transpiration dominates in the dry season and could have substantial effects on moisture recycling patterns (which will be analysed in Part 2). Furthermore, although a change in evaporation partitioning does not change the vegetation interception and transpiration efficiencies, it changes the total evaporation efficiency and the overall temporal distribution of evaporation.

7 Summary and conclusions

This paper developed and evaluated the global hydrological land-surface model STEAM, and used the model output to analyse the terrestrial temporal characteristics of different evaporation fluxes on land. STEAM is designed to 1) be tailored for coupling with the atmospheric moisture recycling
model WAM-2layers, 2) be flexible for land-use change by land-use parametrisation and by including representation of features particularly important for evaporation (e.g., phenology and irrigation), 3) remain simple, transparent and computationally efficient, and 4) simulate evaporation and evaporation partitioning in line with current knowledge. The ability of STEAM to simulate evaporation and evaporation partitioning realistically was evaluated by comparison with other modellings studies, global datasets, and reported values from field studies. STEAM’s total terrestrial evaporation rate (73,900 km$^3$ year$^{-1}$) is comparable with previous estimates – lower than reanalysis products, but higher than other land-surface models. Reasons for this include that we do not add water in data assimilation as in reanalysis, and compared to other land-surface models we use a relatively high precipitation input and also include irrigation and wetlands. Overall, STEAM simulates global evaporation partitioning within the range of previous estimates: 59 % transpiration, 21 % vegetation interception, 10 % floor interception, and 6 % soil moisture evaporation. The global mean transpiration ratio in STEAM is similar to or somewhat higher than other land-surface models, and in line with the recent literature compilation study of [Schlesinger and Jasechko, 2014]. Vegetation interception ratios in STEAM are comparable with both the findings from a global satellite based estimate of interception [Miralles et al., 2010] and with reported values from field studies in the tropics. In agreement with previous studies [McNaughton and Jarvis, 1983; de Bruin and Jacobs, 1989; Teuling et al., 2010], STEAM also simulates higher transpiration ratios in short vegetation types than in forests. Simplifications in STEAM include neglect of runoff routing, groundwater, and sublimation processes. [Koster and Milly, 1997] and [Koster and P. Mahanama, 2012] concluded among others that compatibility between runoff and evaporation formulations can be important due to interaction through soil moisture. Dry season evaporation might also be underestimated by the neglect of groundwater [Miguez-Macho and Fan, 2012] and hydraulic redistribution of soil water by roots [Lee et al., 2005]. Crop simulations presently also do not follow sowing and harvesting dates. The neglect of sublimation can further cause underestimation of interception [Schläpfer et al., 2014]. Nevertheless, the model evaluation analyses and the sensitivity tests suggest that that the current model setup is a reasonable simplification for the research questions asked.

Our analyses show a striking difference in mean annual global time scales between the different evaporation fluxes: 95-434 days for transpiration, 42-46 days for soil moisture evaporation, 5.2-11.6 hours for floor interception, and 1.1-1.6 hours for vegetation interception. The time scales also vary greatly over the seasons and latitudes. Most transpiration occurs several hours or days after a rain event, whereas interception is immediate. We find that 31 % of all transpiration occurs in time steps that have endured more than one day without precipitation, when no vegetation interception occurs. Instead, 96 % of the vegetation interception occurs in a 3 hour time step with precipitation, whereas only 45 % of the transpiration occurs in such conditions. Uncertainties in parametrising storage capacities affect the evaporation partitioning ratios, but have a smaller effect on the relative differences in temporal characteristics. Only the transpiration time scales are significantly changed by changed storage capacity, but are still substantially different from the interception time scales. We note that high vegetation interception ratios coincide with high local evaporation recycling, which suggests that tropical interception may have an important role for vegetation to maintain atmospheric moisture in the air. This will be subject to further investigation in Part 2.

STEAM runs at the same temporal and spatial scale as the atmospheric moisture recycling model WAM-2layers, and can be used in both one and two-way coupling. One-way coupling, i.e., forcing WAM-2layers with STEAM output, is used in Part 2 to investigate the differences in moisture recycling between direct and delayed evaporation fluxes. Two-way coupling, i.e., feeding induced changes in precipitation from WAM-2layers back to STEAM, can be applied in later studies to investigate the effect of land-use change on moisture recycling. Although WAM-2layers does not simulate precipitation, such analyses are possible by assuming that changes in terrestrial evaporation proportionally alters the atmospheric moisture content or the precipitation with continental origin. The importance of land use for the hydrological cycle, the climate, and the Earth system as a whole has been stressed in many studies (e.g., Feddema et al., 2005; Gordon et al., 2005; Rockström et al., 2009a). Thus, changes in evaporative partitioning following e.g., land-use change may have implications and provide answers for landscape resilience, drought development, and effects on remote fresh water resources. The differences in moisture recycling patterns between delayed and direct evaporation fluxes constitutes the case for investigation in Part 2 for the present day situation. Future research should also extend to land-use change scenario analysis to quantify and improve the assessment of land-use change effects on global fresh water resources.
Fig. 1. Water fluxes and stocks in STEAM. Arrows indicate fluxes, and boxes indicate stocks. Dashed lines indicate fluxes and stocks that only exist for particular land-use types. Symbols are listed in Appendix A. A model description is offered in Sect. 2.
Fig. 2. Mean annual evaporation as estimated by STEAM (1999–2008). Grey indicates areas where the evaporative flux is zero. Results are discussed in Sect. 5.1 and 5.2.

Fig. 3. Annual mean STEAM evaporation compared to the statistics (minimum, first quartile, median, third quartile, maximum) of the LandFlux-EVAL product (1989-2005) for (a) merged synthesis, (b) reanalyses, (c) land surface models, and (d) diagnostic datasets. Results are discussed Sect. 5.1.
Fig. 4. Monthly mean STEAM evaporation compared to the statistics (minimum, first quartile, median, third quartile, maximum) of the merged synthesis LandFlux-EV AL product (1989-2005) for (a) January, (b) April, (c) July, and (d) October. Results are discussed Sect. 5.1.

Fig. 5. Partitioned evaporation fluxes expressed as a percentage of total mean annual evaporation (1999–2008). Grey indicates areas where evaporation percentage is zero. Results are discussed in Sect. 5.2.
Fig. 6. Mean monthly evaporation as estimated by STEAM for different latitudes (1999-2008). Note that the scales are different for the different evaporation fluxes. Grey indicates where the evaporative flux is near zero. Results are discussed Sect. 5.2.

Fig. 7. Mean annual runoff of STEAM compared to other datasets (described in Sect. 3). GRDC-Comp (Global Runoff Data Centre composite runoff fields) is the GRDC-WBM (Water Balance Model) runoff corrected using inter-station discharge data. STEAM is run with three settings: with default settings (STEAM: default), with irrigation module switched off (STEAM: no irr) and with 5 % uniform in precipitation forcing (STEAM: 95 % P). STEAM runoff ($P - E - (\Delta S_{\text{snow}}/\Delta t)$) and ERA-I runoff are for the years 1999–2008. GRDC-Comp and GRDC-WBM represent longterm runoff. Results are discussed in Sect. 5.3.
Fig. 8. Comparison between GRDC-Comp (which we consider the benchmark runoff) and the GRDC-WBM, ERA-I, and STEAM runoffs. (a) shows the 1 : 1 agreement line; (b) shows the standard deviations $\sigma$ of GRDC-WBM, ERA-I, and STEAM river basin runoff to GRDC-Comp when Congo is included, and (c) shows the standard deviations when Congo is excluded. Results are discussed in Sect. 5.3.
Fig. 9. Average surface time scales of different evaporation fluxes: (a) transpiration, (b) soil moisture evaporation, (c) vegetation interception, and (d) floor interception (1999–2008). Grey indicates grid cells with mean evaporation rates below 0.01 mm d$^{-1}$. Note that the units are in hours for $E_v$ and $E_t$, and in days for $E_t$ and $E_{sm}$, see Eq. 27. Results are discussed in Sect. 6.1.

Fig. 10. Changes in terrestrial time scales (Eq. 27) over the year and different latitudes (1999-2008). Note that the units are in hours for $E_v$ and $E_t$, and in days for $E_t$ and $E_{sm}$. Grey indicates when time scale approaches infinity. Results are discussed in Sect. 6.1.
Fig. 11. Evaporation partitioning with time since precipitation over terrestrial latitudes (1999-2008). Results are discussed in Sect. 6.2.
### Table 1. Evaporation and evaporation partitioning by land-use type, 1999–2008. Symbols are explained in Appendix A. Results are discussed in Sect. 5.1 and 5.2.

| Land use | Area | \( P \) | \( E \) | \( E_v \) | \( E_t \) | \( E_{sm} \) | \( E_w \) | \( E_v \) of \( E \) | \( E_t \) of \( E \) | \( E_{sm} \) of \( E \) | \( E_w \) of \( E \) | \( E_v \) of \( P \) |
|----------|------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|-------|
| 01: WAT  | 1071 | 937   | 1147  | 0     | 0     | 0     | 0     | 0     | 0     | 0     | 0     | 0     | 100   |
| 02: ENF  | 3224 | 853   | 496   | 155   | 73    | 248   | 20    | 0     | 31    | 50    | 4     | 0     | 18    |
| 03: EBF  | 13541| 2542  | 1208  | 452   | 92    | 652   | 13    | 0     | 37    | 8     | 54    | 1     | 0     |
| 04: DNF  | 1341 | 481   | 366   | 95    | 67    | 191   | 14    | 0     | 26    | 18    | 52    | 4     | 0     |
| 05: DBF  | 1350 | 1057  | 853   | 179   | 83    | 543   | 48    | 0     | 21    | 10    | 64    | 6     | 0     |
| 06: MXF  | 9349 | 958   | 604   | 158   | 38    | 44    | 162   | 0     | 26    | 18    | 52    | 4     | 0     |
| 07: CSH  | 99   | 554   | 499   | 54    | 57    | 324   | 64    | 0     | 11    | 11    | 65    | 13    | 0     |
| 08: OSH  | 21207| 432   | 281   | 38    | 44    | 162   | 37    | 0     | 14    | 16    | 58    | 13    | 0     |
| 09: WSA  | 10585| 1210  | 735   | 103   | 89    | 495   | 48    | 0     | 14    | 12    | 67    | 6     | 0     |
| 10: SAV  | 9904 | 1122  | 861   | 102   | 91    | 602   | 66    | 0     | 12    | 11    | 70    | 8     | 0     |
| 11: GRA  | 18253| 616   | 394   | 54    | 66    | 241   | 33    | 0     | 14    | 17    | 61    | 8     | 0     |
| 12: WET  | 1218 | 1151  | 957   | 114   | 54    | 297   | 513   | 0     | 12    | 3     | 31    | 1     | 54    |
| 13: CRP  | (10352–10851) | 789 | 577   | 99    | 23    | 417   | 38    | 0     | 17    | 4     | 72    | 7     | 13    |
| 14: URB  | 454  | 991   | 465   | 46    | 42    | 256   | 120   | 0     | 10    | 9     | 55    | 26    | 0     |
| 15: MOS  | (7790–7814) | 1262 | 779   | 165   | 79    | 509   | 25    | 0     | 21    | 10    | 65    | 3     | 0     |
| 16: ICE  | 2710 | 560   | 32    | 0     | 32    | 0     | 0     | 0     | 0     | 100   | 0     | 0     |
| 17: BAR  | 18943| 90    | 57    | 1     | 11    | 20    | 25    | 0     | 1     | 19    | 36    | 44    | 0     |
| 18: IRR  | (1060–1195) | 727 | 1375  | 271   | 81    | 910   | 113   | 0     | 20    | 6     | 66    | 8     | 0     |
| 19: RIC  | (175–570) | 1453| 1458  | 242   | 7     | 547   | 4     | 659   | 17    | 0     | 37    | 0     | 45    |

Global 1334688855511558326332459102164413

\( a \) Area varies because a monthly varying irrigation map is applied.

### Table 2. Evaporation of lumped land-use types in comparison with other studies. Results are discussed in Sect. 5.1.

| STEAM, Year 1999–2008 | Gordon et al. (2005) | Schlesinger and Jasechko (2014) | Rockström et al. (1999) based on Mu et al. (2011) |
|-----------------------|----------------------|---------------------------------|-----------------------------------------|
| Area Average \( E_v \) | Area Average \( E_t \) | Area Average \( E_{sm} \) | Area Average \( E_w \) |
| Unit 1000 km\(^2\) | 1000 km\(^2\) | Average \( E \) | Average \( E \) | Average \( E \) |
| Forest\(^a\) | 28805 | 875 | 46665 | 660 |
| Evergreen needleleaf | 3224 | 496 | 2134 | 510 |
| Evergreen broadleaf | 13541 | 1208 | 16278 | 1146 |
| Deciduous needleleaf | 1341 | 366 | 293–795\(^d\) | 458\(^e\) |
| Deciduous broadleaf | 1350 | 853 | 293–795\(^d\) | 549\(^f\) |
| Mixed | 9349 | 604 | 14222 | 313 |
| Savanna | 20489 | 735–861\(^b\) | 19562 | 556 |
| Shrubland | 21306 | 281–499\(^c\) | 18649 | 227 |
| Grassland | 18253 | 393 | 14393 | 258 |

\( a \) Includes all forest types.
\( b \) Woody savannah (09:WSA) and savannah (10:SAV).
\( c \) Closed shrubland (07:CSH) and open shrubland (08:OSH).
\( d \) Deciduous forests in general.
\( e \) Temperate coniferous forest.
\( f \) Temperate deciduous forest.
\( g \) Mediterranean shrubland.
\( h \) Temperate and tropical grassland.
\( i \) Coniferous forest in general.
\( j \) Woody savannah.
\( k \) Wet savannah.
\( l \) Dry shrubland.
\( m \) Cool grassland.
Table 3. Overview of global evaporative partitioning estimates. Results are discussed in Sect. 5.2.

| Unit | $E_t$ | $E_v$ | $(E_t + E_v)$ | Source |
|------|-------|-------|---------------|--------|
| **Land-surface models** |       |       |               |        |
| STEAM | 59    | 21    | 16            | This study |
| JULES (with SiB or SPA scheme) | 38–48 | (Alton et al., 2009) |
| CLM3  | 44    | 17    | 39            | (Lawrence et al., 2007) |
| LPJ   | 65    |       | 36            | (Gerten et al., 2015) |
| A biophysical process-based model | 52    | 20    | 28            | (Choudhury et al., 1998) |
| **Other methods** |       |       |               |        |
| Literature | 61    |       | 53            | (Schlesinger and Jasechko, 2014) |
| Isotope + literature | 35–80 | (Coenders-Gerrits et al., 2014) |
| Isotope + literature | 80–90 | (Jasechko et al., 2013) |
| GLEAM, satellite-based method | 80    | 11    | 7             | (Miralles et al., 2011) |
| Multimodel, GSWP2 | 48    | 16    | 36            | (Dirmeyer et al., 2006) |
Table 4. Comparison of STEAM output (1999–2008) with evaporation and runoff provided by the WaterMIP (Water Model Intercomparison Project) (1985–1999) ([Haddeland et al., 2011] [Harding et al., 2011]). The ERA-I precipitation used to force STEAM and the WFD (Watch Forcing Data) precipitation used to force WaterMIP are also shown for each compared river basin. Results are discussed in Sect. 5.3.

| Unit       | \( E_{\text{STEAM}} \) | \( E_{\text{WaterMIP}} \) | \( Q_{\text{STEAM}} \) | \( Q_{\text{WaterMIP}} \) | \( P_{\text{ERA-I}} \) | \( P_{\text{WFD}} \) |
|------------|----------------|----------------|----------------|----------------|----------------|----------------|
|           | Low | Mean | High | Low | Mean | High | Low | Mean | High | Low | Mean | High |
| Amazon     | 1154 | 1021 | 1195 | 1430 | 1228 | 815 | 1043 | 1207 | 2382 | 2243 |      |      |
| Mississippi| 595 | 492  | 642  | 747 | 93   | 167 | 269  | 418  | 692  | 909  |      |      |
| Ganges/Brahmaputra | 739 | 410  | 546  | 828 | 809  | 553 | 891  | 1038 | 1555 | 1447 |      |      |
| Lena       | 319 | 172  | 230  | 283 | 148  | 103 | 151  | 211  | 487  | 385  |      |      |
| Global     | 555 | 415  | 499  | 586 | 325  | 290 | 375  | 457  | 888  | 872  |      |      |
Table 5. Robustness to storage capacity parametrisation of STEAM, (global mean for 1999–2008). The subscript \( t \) stands for transpiration, \( sm \) for soil moisture evaporation, \( v \) for vegetation interception, \( f \) for floor interception, and \( uz \) for unsaturated zone. Methods are described in Sect. 4.2.3 and results are discussed in Sect. 6.1 and 6.2.

| Storage capacity          | Default | Transpiration-plus | Interception-plus |
|---------------------------|---------|--------------------|-------------------|
| \( S_{v,\text{max}} \)   | 100 %   | 50 %               | 150 %             |
| \( S_{t,\text{max}} \)   | 100 %   | 50 %               | 150 %             |
| \( S_{uz,\text{max}} \)  | 100 %   | 120 %              | 80 %              |
| **Total evaporation**     | 73,900 km\(^3\) year\(^{-1}\) | 73,200 km\(^3\) year\(^{-1}\) | 74,200 km\(^3\) year\(^{-1}\) |
| **Evaporation ratio**     |         |                    |                   |
| \( E_t / E \)             | 59 %    | 64 %               | 54 %              |
| \( E_{sm} / E \)          | 6 %     | 7 %                | 5 %               |
| \( E_v / E \)             | 21 %    | 12 %               | 27 %              |
| \( E_f / E \)             | 10 %    | 12 %               | 10 %              |
| **Time scales**           |         |                    |                   |
| \( \tau_{ts,t} \)         | 274 days| 434 days           | 95 days           |
| \( \tau_{ts,sm} \)        | 42 days | 43 days            | 46 days           |
| \( \tau_{ts,v} \)         | 1.3 hours | 1.1 hours         | 1.6 hours         |
| \( \tau_{ts,f} \)         | 7.7 hours | 5.2 hours         | 11.6 hours        |
| **Evaporation efficiency,**|         |                    |                   |
| \( \beta_{\text{wet}} \), 3 hours after precipitation\(^a\) | 58 %    | 56 %               | 60 %              |
| \( \beta_{\text{wet},t} \) | 45 %    | 46 %               | 43 %              |
| \( \beta_{\text{wet},sm} \) | 39 %    | 44 %               | 35 %              |
| \( \beta_{\text{wet},v} \) | 96 %    | 96 %               | 96 %              |
| \( \beta_{\text{wet},f} \) | 83 %    | 87 %               | 79 %              |
| **Evaporation efficiency,**|         |                    |                   |
| \( \beta_{\text{dry}} \), 24 hours without precipitation\(^b\) | 23 %    | 24 %               | 21 %              |
| \( \beta_{\text{dry},t} \) | 31 %    | 31 %               | 31 %              |
| \( \beta_{\text{dry},sm} \) | 32 %    | 29 %               | 34 %              |
| \( \beta_{\text{dry},v} \) | 1 %     | 1.2 %              | 0.8 %             |
| \( \beta_{\text{dry},f} \) | 3.9 %   | 2.8 %              | 5 %               |

\(^a\) The evaporation efficiency is calculated for 3 hour time steps with precipitation.

\(^b\) The evaporation efficiency is calculated for 3 hour time steps that have been without precipitation for more than 24 hours.
Appendix A

Notations

Symbols used in this paper are listed and defined in Table A1.
Table A1. List of symbols.

| Symbol | Units | Description |
|--------|-------|-------------|
| $\alpha$ | – | Albedo |
| $\beta$ | – | Evaporation efficiency, i.e., the portion of evaporation evaporated during certain conditions. |
| $\gamma$ | kPa K$^{-1}$ | Psychrometric constant |
| $\Delta n$ | h | Time step, 24 h |
| $\Delta t$ | h | Time step, 3 h |
| $\delta$ | kPa K$^{-1}$ | Slope of the saturated vapour pressure curve |
| $\eta_{\text{clay}}$ | % | Clay content of the top soil |
| $\Theta_{\text{top}}$ | – | Effective saturation of top soil |
| $\theta_{\text{top}}$ | – | Volumetric soil moisture content of top soil |
| $\theta_{\text{top,sat}}$ | – | Volumetric soil moisture content of top soil at saturation |
| $\theta_{\text{top,res}}$ | – | Volumetric soil moisture content of top soil at residual point |
| $\theta_{\text{sat}}$ | – | Volumetric soil moisture content of the unsaturated zone |
| $\theta_{\text{uz,fc}}$ | – | Volumetric soil moisture content at field capacity in the unsaturated zone |
| $\theta_{\text{uz,wp}}$ | – | Volumetric soil moisture content at wilting point |
| $\kappa$ | – | Von Kármán constant, 0.41. |
| $\lambda$ | MJ kg$^{-1}$ | Latent heat of vaporisation of water |
| $\xi_{\text{mw}}$ | – | Ratio of the molecular weight of water vapour to that for dry air, 0.622. |
| $\rho_a$ | kg m$^{-3}$ | Density of air. |
| $\rho_w$ | kg m$^{-3}$ | Density of water |
| $\tau_s$ | day | Mean terrestrial time scale |
| $\phi_{\text{lu}}$ | – | Land-use fraction |
| $\phi_{\text{ow}}$ | – | Open water fraction |
| $\phi_{\text{vs}}$ | – | Vegetation in soil fraction |
| $\phi_{\text{vw}}$ | – | Vegetation in water fraction |
| $\chi$ | h | Top soil moisture dry out time parameter |
| $\chi_{\text{min}}$ | h | Minimum top soil moisture dry out time parameter, 60 h. |
| $C_p$ | MJ kg$^{-1}$ K$^{-1}$ | Heat capacity of water at constant pressure, $1.01 \times 10^{-3}$ MJ kg$^{-1}$ K$^{-1}$ |
Table A1. Continued.

| Symbol  | Units | Description                                                                 |
|---------|-------|------------------------------------------------------------------------------|
| $c_{AR}$ |       | Area reduction factor, 0.4                                                     |
| $c_{D1}$ |       | Vapor pressure stress parameter, 3.                                            |
| $c_{D2}$ |       | Vapor pressure stress parameter, 0.1.                                         |
| $c_{R}$  |       | Radiation stress parameter, 100.                                              |
| $c_{sc}$ |       | Storage capacity factor, 0.2.                                                  |
| $c_{uz}$ |       | Soil moisture stress parameter, 0.07                                           |
| $D_{0.5}$ | kPa   | Vapour pressure deficit coefficient, 1.5 kPa                                  |
| $D_a$    | kPa   | Vapour pressure deficit                                                       |
| $d$      | m     | Zero plane displacement                                                       |
| $E$      | m d$^{-1}$ | Total evaporation                                                           |
| $E_t$    | m d$^{-1}$ | Floor interception evaporation                                               |
| $E_{t,lu}$ | m (Δt)$^{-1}$ | Land-use specific floor interception evaporation in $\phi_{vs}$         |
| $E_p$    | m (Δt)$^{-1}$ | Potential evaporation                                                         |
| $E_{p,day}$ | m d$^{-1}$ | Potential evaporation                                                         |
| $E_{am}$ | m d$^{-1}$ | Soil moisture evaporation                                                     |
| $E_{am,lu}$ | m (Δt)$^{-1}$ | Land-use specific soil moisture evaporation in $\phi_{vs}$          |
| $E_t$    | m d$^{-1}$ | Transpiration evaporation                                                     |
| $E_{t,lu}$ | m (Δt)$^{-1}$ | Land-use specific transpiration in $\phi_{vs}$                               |
| $E_{t,vw}$ | m (Δt)$^{-1}$ | Land-use specific transpiration in $\phi_{vw}$                               |
| $E_v$    | m d$^{-1}$ | Vegetation interception evaporation                                           |
| $E_{v,lu}$ | m (Δt)$^{-1}$ | Land-use specific vegetation interception evaporation in $\phi_{vs}$|
| $E_{v,lu}$ | m (Δt)$^{-1}$ | Land-use specific vegetation interception evaporation in $\phi_{vw}$|
| $E_{v,ow}$ | m d$^{-1}$ | Open water evaporation                                                        |
| $E_{v,ow}$ | m (Δt)$^{-1}$ | Land-use specific water evaporation in $\phi_{ow}$                          |
| $E_{v,ow}$ | m (Δt)$^{-1}$ | Land-use specific open water evaporation in $\phi_{ow}$                |
| $e_a$    | kPa   | Actual vapor pressure                                                        |
| $e_s$    | kPa   | Saturated vapor pressure                                                     |
| $G$      | MJ m$^{-2}$ d$^{-1}$ | Ground heat flux                                                                |
| $h$      | m     | Plant height                                                                 |
| $h_{max}$ | m     | Minimum plant height                                                         |
### Table A1. Continued.

| Symbol | Units | Description |
|--------|-------|-------------|
| $h_{\text{min}}$ | m | Maximum plant height |
| $I_t$ | md$^{-1}$ | Irrigation applied to $S_t$ |
| $I_g$ | md$^{-1}$ | Gross irrigation |
| $I_{\text{req}}$ | m ($\Delta t$)$^{-1}$ | Irrigation requirement |
| $I_{az}$ | md$^{-1}$ | Irrigation applied to $S_{az}$ |
| $I_v$ | md$^{-1}$ | Irrigation applied to $S_v$ |
| $i_{GS}$ | – | Growing Season Index |
| $i_{LA}$ | m$^2$m$^{-2}$ | Leaf Area Index |
| $i_{LA,\text{eff}}$ | m$^2$m$^{-2}$ | Effective Leaf Area Index |
| $i_{LA,\text{max}}$ | m$^2$m$^{-2}$ | Maximum Leaf Area Index |
| $i_{LA,\text{min}}$ | m$^2$m$^{-2}$ | Minimum Leaf Area Index |
| $J_{\text{add}}$ | m ($\Delta t$)$^{-1}$ | Water added in water stores to compensate for lack of horizontal flows |
| $k$ | – | Function of $r_a$ and $r_s$ |
| $N$ | s | Day length |
| $N_{\text{high}}$ | s | Day length, higher sub-optimal threshold, assumed to be 39 600 s. |
| $N_{\text{low}}$ | s | Day length, lower sub-optimal threshold, assumed to be 36 000 s. |
| $P$ | md$^{-1}$ | Total precipitation |
| $P_{\text{eff}}$ | md$^{-1}$ | Effective precipitation, (i.e., overflow from floor interception stock to unsaturated zone stock) |
| $P_{\text{melt}}$ | md$^{-1}$ | Snowmelt |
| $P_t$ | m ($\Delta t$)$^{-1}$ | Rainfall |
| $P_d$ | m ($\Delta t$)$^{-1}$ | Snowfall |
| $P_{\text{tf}}$ | m ($\Delta t$)$^{-1}$ | Throughfall, (i.e., overflow from vegetation interception stock to floor interception stock) |
| $p$ | kPa | Atmospheric pressure |
| $Q_{az}$ | m ($\Delta t$)$^{-1}$ | Outflow from $S_{az}$ |
| $Q_w$ | m ($\Delta t$)$^{-1}$ | Runoff from $S_w$ |
| $R_{\text{net}}$ | MJ m$^{-2}$ d$^{-1}$ | Net radiation |
| $R_{\text{net, lw}}$ | MJ m$^{-2}$ d$^{-1}$ | Net long wave radiation |
| $R_{\text{sw}}$ | MJ m$^{-2}$ d$^{-1}$ | Short wave radiation |
Table A1. Continued.

| Symbol       | Units        | Description                                           |
|--------------|--------------|-------------------------------------------------------|
| $r_a$        | dm$^{-1}$    | Aerodynamic resistance                                |
| $r_{a,f}$    | dm$^{-1}$    | Floor aerodynamic resistance                          |
| $r_{a,v}$    | dm$^{-1}$    | Vegetation aerodynamic resistance                     |
| $r_{a,w}$    | dm$^{-1}$    | Open water aerodynamic resistance                     |
| $r_s$        | dm$^{-1}$    | Surface resistance                                    |
| $r_{s,sm}$   | dm$^{-1}$    | Surface soil moisture resistance                      |
| $r_{s,sm,min}$| dm$^{-1}$    | Minimum surface soil moisture resistance              |
| $r_{s,st}$   | dm$^{-1}$    | Surface stomatal resistance                           |
| $r_{s,st,min}$| dm$^{-1}$    | Minimum surface stomatal resistance                   |
| $S_f$        | m            | Floor interception stock                              |
| $S_{f,lu}$   | m            | Floor interception stock of a specific land-use type  |
| $S_{f,ma}$   | m            | Floor interception storage capacity                    |
| $S_{snow}$   | m            | Snow stock                                            |
| $S_{us}$     | m            | Unsaturated stock                                     |
| $S_{us,lu}$  | m            | Unsaturated stock of a specific land-use type          |
| $S_{us,ma}$  | m            | Unsaturated storage capacity                          |
| $S_{us,sm}$  | m            | Unsaturated stock available for soil moisture evaporation|
| $S_{us,t}$   | m            | Unsaturated stock available for transpiration         |
| $S_v$        | m            | Vegetation interception stock                         |
| $S_{v,lu}$   | m            | Vegetation interception stock of a specific land-use type|
| $S_{v,ma}$   | m            | Vegetation interception storage capacity              |
| $S_w$        | m            | Water stock                                           |
| $S_{w,lu}$   | m            | Water stock of a specific land-use type               |
### Table A1. Continued.

| Symbol | Units | Description |
|--------|-------|-------------|
| $T_{\text{dew}}$ | K | Dew point temperature |
| $T_{\text{mean}}$ | K | Daily mean temperature |
| $T_{\text{min}}$ | K | Daily minimum temperature |
| $T_{\text{min, high}}$ | K | Daily minimum temperature, higher sub-optimal threshold, 278.15 K. |
| $T_{\text{min, low}}$ | K | Daily minimum temperature, lower sub-optimal threshold, 271.15 K. |
| $T_{\text{opt}}$ | K | Optimum photosynthesis temperature |
| $u_{10}$ | m·d$^{-1}$ | Wind speed at 10 m height |
| $u_{200}$ | m·d$^{-1}$ | Wind speed at 200 m height |
| $u_{\text{ref}}$ | m·d$^{-1}$ | Wind speed at reference height |
| $y_{\text{u}}$ | m | Depth of the unsaturated zone |
| $y_{\text{top}}$ | m | Depth of the top soil |
| $Z$ | m | Elevation |
| $z_0$ | m | Aerodynamic roughness length |
| $z_{0,f}$ | m | Roughness length of substrate floor |
| $z_{10}$ | m | Height of wind speed $u_{10}$ |
| $z_{200}$ | m | Height of wind speed $u_{200}$ |
| $z_{\text{ref}}$ | m | Reference height |
Appendix B

Model equations

B1  Input variables to the Penman–Monteith equation

The vapour pressure deficit $D_a$ is defined as:

$$D_a = e_s - e_a(T_{dew}),$$  \hspace{1cm} (B1)

where $e_s$ [kPa] is the saturated vapour pressure at temperature $T_{\text{mean}}$ [K] and estimated from the average of the saturated vapour pressures of the daily maximum and minimum temperature, $e_a$ [kPa] is the vapour pressure of air at height $z_{\text{ref}}$ [m], and $T_{dew}$ [K] is the daily mean dew point temperature.

Vapor pressure $e_a$ is estimated from the formula below:

$$e_a(T_{dew}) = \frac{0.6108 e_a^{17.27(T_{dew} - 273.15)}}{T_{dew} - 35.85}.$$  \hspace{1cm} (B2)

For the estimation of $e_s$, $T_{dew}$ was replaced by $T_{\text{max}}$ or $T_{\text{min}}$. The latent heat of water vapourisation $\lambda$ [MJkg$^{-1}$] is expressed as:

$$\lambda = 2.501 - 0.002361(T_{\text{mean}} - 273.15).$$  \hspace{1cm} (B3)

The gradient $\delta$ [kPa.K$^{-1}$] of the saturated vapour pressure function is given by

$$\delta = \frac{4098 \times e_s}{237.3 + (T_{\text{mean}} - 273.15)^2}.$$  \hspace{1cm} (B4)

The psychrometric constant $\gamma$ [kPaK$^{-1}$] is

$$\gamma = \frac{C_p p}{\xi_{\text{sw}} \lambda},$$  \hspace{1cm} (B5)

where $p$ is the atmospheric pressure [kPa], and $\xi_{\text{sw}}$ is the ratio of the molecular weight of water vapour to that for dry air [0.622].

Net radiation is calculated by:

$$R_{\text{net}} = (1 - \alpha) R_{\text{sw}} - R_{\text{net, lw}}$$  \hspace{1cm} (B6)

where $\alpha$ is albedo, $R_{\text{sw}}$ is the incoming shortwave radiation and $R_{\text{net, lw}}$ is the outgoing net longwave radiation. In reality, albedo varies with angle of reflection and the surface properties such as snow cover change and soil wetness. Here, we assume $\alpha$ to be fixed for each land-use type, see Table C1.

Daily ground heat flux $G$ is derived from interpolating monthly ground heat flux $G_{\text{month}}$ (Allen et al., 1998)

$$G_{\text{month}} = 0.07(T_{\text{month} +1 } - T_{\text{month} -1}).$$  \hspace{1cm} (B7)

There are three types of aerodynamic resistances used in STEAM: the aerodynamic vegetation resistance $r_{a,v}$, the aerodynamic floor resistance $r_{a,f}$, and the aerodynamic water resistance $r_{a,w}$. They are expressed as follows (Shuttleworth 2012):

$$r_{a,v} = \frac{\ln \left( \frac{z_{\text{ref}, v}}{z_0} - d \right) \ln \left( \frac{z_{\text{ref}, v}}{z_0} - d \right) \left( \frac{200}{z_0} \right)}{u_{\text{ref}, v} R^{2}},$$  \hspace{1cm} (B8)

$$r_{a,f} = \frac{\ln \left( \frac{z_{\text{ref}, f}}{z_0} - d \right) \ln \left( \frac{z_{\text{ref}, f}}{z_0} - d \right) \left( \frac{200}{z_0} \right)}{u_{\text{ref}, f} R^{2}},$$  \hspace{1cm} (B9)

$$r_{a,w} = \frac{4.72h^{3/2} \left( \frac{200}{z_0} \right) \left( \frac{200}{z_0} \right)}{1 + 0.536 u_{\text{ref}, w} R},$$  \hspace{1cm} (B10)

where $z_{\text{ref}}$ is the reference height [m], $z_0$ is the aerodynamic roughness length [m], $d$ is the zero-plane displacement height [m] and $u_{\text{ref}}$ is the wind speed [m] at $z_{\text{ref}}$. Wind speed $u_{\text{ref}}$ is estimated from wind speed $u_{10}$ given by ERA-1 at 10 m $z_{10}$ [m] under the assumption of a logarithmic wind profile and stable neutral atmospheric conditions:

$$u_{\text{ref}, f} = \frac{\ln \left( \frac{z_{\text{ref}, f}}{z_0} - d \right) \ln \left( \frac{200}{z_0} \right)}{\ln \left( \frac{200}{z_0} \right)}.$$  \hspace{1cm} (B11)

$$u_{\text{ref}, w} = \frac{\ln \left( \frac{z_{\text{ref}, w}}{z_0} - d \right) \ln \left( \frac{200}{z_0} \right)}{\ln \left( \frac{200}{z_0} \right)}.$$  \hspace{1cm} (B12)

where the reference height $z_{\text{ref}, f}$ and $z_{\text{ref}, w}$ are 2 m and $z_{\text{ref}, v}$ is $2 + h$ [m], with $h$ being the plant height [m]. However, because some vegetation is higher than 10 m, wind speed at 200 m is substituted into the formula to derive wind speeds at lower elevations:

$$u_{\text{ref}, v} = \frac{\ln \left( \frac{z_{\text{ref}, v}}{z_0} - d \right) \ln \left( \frac{200}{z_0} \right)}{\ln \left( \frac{200}{z_0} \right)}.$$  \hspace{1cm} (B13)

The aerodynamic roughness length $z_0$ [m] is estimated from:

$$z_0 = \begin{cases} z_{0,f} + 0.29h \sqrt{0.2 i_{LA}} & i_{LA} \leq 1, \\ 0.3h (1 - d/h) & i_{LA} > 1. \end{cases}$$  \hspace{1cm} (B14)

Zero plane displacement $d$ is estimated from $h$ [m] and $i_{LA}$ [m$^2$m$^{-2}$]

$$d = 1.1h \ln \left[ 1 + (0.2 i_{LA})^{0.25} \right].$$  \hspace{1cm} (B15)

B2  Surface stomatal resistance

Surface resistance applies only to transpiration and soil moisture evaporation, since interception and open water evaporation occur without resistance. The surface stomatal resistance $r_{s, st}$ of vegetation is simulated by the Jarvis–Stewart equation (Stewart, 1988), taking into account of solar radiation, vapour pressure deficit, optimum temperature, and soil moisture stress:

$$r_{s, st} = \frac{r_{s, st, min}}{i_{LA, eff} f (R_{\text{sw}}) f (D_a) f (T_{\text{mean}}) f (\theta_u)},$$  \hspace{1cm} (B17)
where $r_{s,t}$, min is the minimum surface stomatal resistance dependent on land-use type and specified in the land-use look-up table (Table C1). $i_{LA,eff}$ is the effective leaf area index (unit leaf area per unit ground area that is actively participating in transpiration) and $f$ are the four stress functions for incoming shortwave radiation $R_{sw}$ in W m$^{-2}$, vapour pressure deficit $D_a$, mean daily temperature $T_{mean}$ and soil moisture $\theta_{uz}$ (Stewart, 1988). Effective leaf area index $i_{LA,eff}$ is adapted from Allen et al. (2006) and Zhou et al. (2006) as:

$$i_{LA,eff} = \frac{i_{LA}}{0.2i_{LA} + 1}.$$  \hfill (B18)

The stress functions vary between 0 and 1. The stress function of soil moisture $f(\theta_{uz})$ is the same as in Eq. (19). The other stress functions as follows (Jarvis, 1976; Zhou et al., 2006; Matsumoto et al., 2008):

$$f(R_{sw}) = R_{sw} \left(1 + c_R/1000\right) \left(c_R + R_{sw}\right)^{-1}, \quad (B19)$$

$$f(D_a) = \left[1 + \left(D_a/D_{0.5}\right)^{cD1}\right]^{-1} \left(1 - cD2\right) + cD2, \quad (B20)$$

$$f(T_{mean}) = \begin{cases} 0 & T_{mean} < 273.15 \\ 1 - T_{opt}^{-2}(T_{mean} - T_{opt})^2 & (T_{mean} > T_{opt} + 1) \\ \cup (273.15 \leq T_{mean} < T_{opt} - 1), & T_{opt} - 1 \leq T_{mean} \leq T_{opt} + 1 \\ 1 & \end{cases} \quad (B21)$$

where $c_R$ is the radiation stress parameter fixed at 100 (Zhou et al., 2006), $D_{0.5}$ is the vapour pressure deficit halfway between 1 and $cD2$ set at 1.5 kPa, $cD1$ is the first vapour pressure parameter set at 3, and $cD2$ is the second vapour pressure stress parameter set at 0.1 (Matsumoto et al., 2008). Optimum temperature $T_{opt}$ [K] is based on elevation a.s.l. $Z$ [m] and latitude $\omega$ [rad] (Cui et al., 2012):

$$T_{opt} = 302.45 - 0.003(Z - |\omega|). \quad (B22)$$

Graphical representations of the stress functions are presented in Fig. B1. Under unfavourable conditions where at least one of the stress functions equals zero, $r_{s,t}$ is assumed to be 0.58 dm$^{-1}$ (50 000 sm$^{-1}$), corresponding to the molecular diffusivity of water vapour through leaf cuticula (Tourula and Heikinheimo, 1998). If $i_{LA}$ is zero, no transpiration is allowed.
Appendix C

Primary land-use parameters

The parameters used to describe land use include maximum and minimum leaf area index $i_{LA,\text{max}}$ and $i_{LA,\text{min}}$, maximum and minimum plant height $h_{\text{max}}$ and $h_{\text{min}}$, depth of the unsaturated zone (or rather active rooting depth) $y_{uz}$, albedo $\alpha$, minimum stomatal resistance $r_{s,\text{st, min}}$ and floor roughness $z_{0,f}$. Land-use parameters considered include those used in other large-scale land-surface or hydrological models (Federer et al., 1996; van den Hurk et al., 2000; van den Hurk, 2003; Zhou et al., 2006; Bastiaanssen et al., 2012), and studies of specific land-use properties (Scurlock et al., 2001; Zeng, 2001; Breuer et al., 2003; Kleidon, 2004). The range of parameters in the literature can sometimes be significant and contradictory, due to discrepancies in scale, parameter definitions, and methods of parameter estimation. The choice of land-use parameters is therefore not simply taken as a mean from the literature values investigated, but rather based on the preservation of the internal consistency of STEAM, manual calibration and priority for literature values with higher relevance. In addition, some land-use types are assumed to contain water, either as water below vegetation or as open water. The land-use parameters used in the model are shown in Table C1, and the parametrisation of water fractions are presented in Table C2.
Fig. B1. Stress functions used in the Jarvis–Stewart equation (See Eq. B16.).

Table C1. Land-use parameters used in STEAM. For model description, see Sect. 2.

| Land-use class     | \( r_{A,\text{max}} \) | \( r_{A,\text{min}} \) | \( y_{ulc} \) | \( \alpha \) | \( h_{\text{max}} \) | \( h_{\text{min}} \) | \( z_{0,f} \) | \( r_{s, \text{st, min}} \) |
|--------------------|------------------------|------------------------|---------------|--------------|----------------|----------------|---------------|----------------|
| 01: WAT (Water)    | --                     | --                     | m             | m            | m             | m             | 0.00137       | 0              |
| 02: ENF (Evergreen needleleaf forest) | 5.5                     | 2                       | 2             | 0.15         | 17            | 17            | 0.02          | 300            |
| 03: EBF (Evergreen broadleaf forest) | 5.5                     | 2                       | 2             | 0.18         | 30            | 30            | 0.02          | 200            |
| 04: DNF (Deciduous needleleaf forest) | 5                       | 1                       | 2             | 0.18         | 17            | 17            | 0.02          | 300            |
| 05: DBF (Deciduous broadleaf forest) | 5.5                     | 1                       | 2             | 0.18         | 25            | 25            | 0.02          | 200            |
| 06: MXF (Mixed forest) | 5                       | 1                       | 2             | 0.18         | 20            | 20            | 0.02          | 250            |
| 07: CSH (Closed shrubland) | 1.5                     | 0.5                     | 2             | 0.2          | 0.8           | 0.8           | 0.02          | 200            |
| 08: OSH (Open shrubland) | 1.5                     | 0.5                     | 2             | 0.2          | 1             | 1             | 0.02          | 200            |
| 09: WSA (Woody savannah) | 2                       | 0.5                     | 2             | 0.2          | 0.8           | 0.8           | 0.02          | 150            |
| 10: SAV (Savannah)    | 2                       | 0.5                     | 3.5           | 0.2          | 0.8           | 0.1           | 0.02          | 150            |
| 11: GRA (Grassland)   | 2                       | 0.5                     | 1.5           | 0.2          | 0.8           | 0.8           | 0.05          | 0.01           |
| 12: WET (Permanent wetland) | 4                       | 1                       | 1.5           | 0.15         | 1             | 0.05          | 0.01          | 150            |
| 13: CRP (Cropland, rainfed) | 3.5                     | 0.5                     | 1.5           | 0.2          | 0.8           | 0.8           | 0.05          | 0.005          |
| 14: URB (Urban and built-up) | 1                       | 0.1                     | 0.5           | 0.18         | 0.8           | 0.8           | 0.001         | 250            |
| 15: MOS (Crop/natural mosaic) | 3.5                     | 0.5                     | 1.5           | 0.2          | 0.8           | 0.1           | 0.005         | 150            |
| 16: ICE (Snow/ice)    | 0                       | 0                       | 0             | 0.7          | 0             | 0             | 0.001         | 0              |
| 17: BAR (Barren land) | 0.1                     | 0.01                    | 1.5           | 0.25         | 0.8           | 0             | 0.001         | 200            |
| 18: IRR (Irrigated crop, excl. rice) | 3.5                     | 3.5                     | 0.5           | 0.2          | 0.8           | 0.8           | 0.005         | 150            |
| 19: RIC (Irrigated rice paddies) | 3.5                     | 3.5                     | 0.5           | 0.2          | 0.8           | 0.8           | 0.005         | 150            |

* The unit for \( r_{s, \text{st, min}} \) is \( d m^{-1} \) throughout the paper, and only given as \( m^{-1} \) in this table to facilitate comparison with other studies.
Table C2. Fractions of vegetation in soil $\phi_{vs}$, vegetation in water $\phi_{vw}$, and open water $\phi_{ow}$ by land-use type. Related equations are described in Sect. 2.2.

| Land-use type | $\phi_{vs}$ | $\phi_{vw}$ | $\phi_{ow}$ |
|---------------|-------------|-------------|-------------|
| 12:WET        | 1/3         | 1/3         | 1/3         |
| 19:RIC        | 1/10        | 9/10        | 0           |
| 01:WAT        | 0           | 0           | 1           |
| Other         | 1           | 0           | 0           |
Appendix D

Sensitivity to precipitation

We perform a sensitivity check against precipitation because STEAM is forced by ERA-I precipitation reanalyses data, which is higher than several other satellite and/or gauge-based precipitation datasets. For the 1999–2008, the mean global ERA-I precipitation is 118 236 km$^3$ year$^{-1}$ for a land area of 133 146 465 km$^2$. Other reported terrestrial precipitation values include 111 000 km$^3$ year$^{-1}$ (Oki and Kanae, 2006), 109 500 km$^3$ year$^{-1}$ from CRU, 111 200 km$^3$ year$^{-1}$ from PREC/L, and 112 600 km$^3$ year$^{-1}$ from GPCP (Trenberth et al., 2007).

Table D1 provides an overview of the sensitivity of runoff and evaporation fluxes to a uniform 5% reduction in precipitation. A number of observations can be noted. First, the mean annual STEAM runoff is clearly more sensitive (−10.95%) to precipitation reduction compared to evaporation (−1.78%). Second, among the evaporation fluxes, soil moisture evaporation (−2.95%) and transpiration (−2.32%) respond most strongly, whereas the vegetation (−0.89%) and floor interception (−0.65%) evaporation fluxes reduce only marginally. This is logical, because interception stocks are already small and depend more on rainfall frequency than on rainfall amount. Third, the increase in open water evaporation (+0.25%) is small, and can be explained by decreases in vegetation interception that translated into increases in available energy for water evaporation in wetlands and rice paddies. Fourth, the relative reduction in snow accumulation (−14.63%) is high since snow melt is unchanged. Last, the global mean evaporative partitioning is changed only insignificantly towards lower transpiration ratio.

The sensitivity of transpiration is highest over the US, Australia, the subtropical South America and Africa, and other areas that at least during part of the years are water constrained. In the wet tropics, transpiration rates do not react to precipitation reductions. Vegetation interception experiences an insignificant relative decrease, which is highest in the north and highest in the tropics. This is probably caused by a combination of lower original interception rates in the boreal forests, and the relatively higher dependence on high rainfall frequency in the tropical forests.

This uniform perturbation of precipitation forcing indicates that STEAM evaporation is much less sensitive to precipitation than runoff. This can be explained by the fact that evaporation is constrained by potential evaporation, which relates to other factors than just precipitation. In wet regions where soil moisture is close to saturation, any excess precipitation would more likely lead to increase in runoff rather than evaporation. The sensitivity of runoff to precipitation data is also reported in the literature (e.g., Fekete et al., 2004; Materia et al., 2010) and supports the view that runoff comparisons will not accurately describe how well land-surface models estimate evaporation when precipitation is uncertain.
Table D1. Overview of the sensitivity of runoff, evaporation, and model snow accumulation to uniform reduction in precipitation quantity, (global mean for 1999–2008).

| Flux     | Default km$^3$ year$^{-1}$ | % $E$ | 5 % reduction in $P$ km$^3$ year$^{-1}$ | % $E$ | Change % |
|----------|-----------------------------|-------|----------------------------------------|-------|----------|
| $P$      | 118 236                     | –     | 112 324                                | –     | –5       |
| $Q$      | 43 216                      | –     | 38 762                                 | –     | –10.3    |
| $E_t$    | 73 933                      | 100   | 72 644                                 | 100   | –1.74    |
| $E_{et}$ | 43 376                      | 58.7  | 42 392                                 | 58.4  | –2.27    |
| $E_v$    | 15 288                      | 20.7  | 15 152                                 | 20.9  | –0.89    |
| $E_t$    | 7706                        | 10.4  | 7 657                                  | 10.5  | –0.64    |
| $E_{soil}$ | 4335                        | 5.9   | 4 207                                  | 5.8   | –2.95    |
| $E_{sw}$ | 3228                        | 4.4   | 3236                                  | 4.5   | +0.25    |
| $\text{d}S_{\text{snow}}/\text{dt}$ | 1087 | – | 918 | – | –15.5 |
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