



Hydroclimate responses to forest cover change

Mechanistic insights into tropical circulation and hydroclimate responses to future forest cover change

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Abstract. Afforestation and the prevention of deforestation are important climate mitigation strategies, alongside reductions in greenhouse gas emissions. However, the biogeophysical effects of future forest cover change on the atmospheric circulation and tropical hydroclimate remain uncertain. We address this research gap using future scenario simulations from seven multi-ensemble models participating in the Land Use Model Intercomparison Project (LUMIP). The largest afforestation and avoided deforestation areas are located in the tropics, leading to robust increases in local evapotranspiration and precipitation, but widespread decreases in net precipitation (precipitation minus evapotranspiration), especially over Africa. Our results suggest that two competing mechanisms shape the tropospheric circulation and net precipitation response over Africa: The increased surface roughness not only increases evaporation, but also surface momentum fluxes, thereby slowing near-surface winds and reducing the orographic net precipitation. Opposing this surface drag effect is an energetic effect due to increased net energy input to the atmosphere, which strengthens convection and increases net precipitation. While the surface drag effect dominates and leads to a net precipitation decrease over western and southern Africa, the energetic effect dominates and leads to a net precipitation increase over central Africa. This tropical hydroclimate response to the forest cover change is largely independent of the background climate under low- to medium-warming scenarios. Our findings contribute to an improved understanding of the mechanisms of forest cover impact on future hydroclimate changes in the tropics and highlight the importance of considering hydroclimatic feedbacks in the context of future afforestation strategies.

Short summary

Afforestation is a key strategy for climate change mitigation, yet the impacts on tropical hydroclimate remain uncertain. We find that future afforestation would increase evaporation and precipitation in the tropics, especially over Africa. It would also reduce net precipitation (precipitation minus evaporation) in these regions, which determines water availability. This happens because trees slow near-surface winds, while their influence on the energy budget strengthens convection.



1 Introduction

Afforestation and the prevention of deforestation are widely recognized as key strategies for climate change mitigation by enhancing carbon sequestration and strengthening the terrestrial carbon sink, alongside efforts to reduce greenhouse gas emissions (Roe et al., 2021; Girardin et al., 2021; Seddon et al., 2021). All scenarios in the IPCC AR6 report that limit warming to 2°C or less by 2100 rely on carbon dioxide removal (CDR) in addition to emissions reductions (IPCC, 2022). Nature-based solutions such as afforestation are increasingly valued for their cost-effectiveness compared to technological CDR methods, like direct air carbon capture (DACCS) (Griscom et al., 2017). Moreover, forest expansion and conservation provide additional co-benefits, including biodiversity preservation, improved water quality, and enhanced societal well-being (Chausson et al., 2020; Seddon, 2022).

The cooling due to forests' influence on the carbon and other biogeochemical cycles can be either enhanced or partially to fully offset by biogeophysical effects—climate responses to changes in surface albedo, evaporative efficiency, and surface roughness—depending on the location of forestation (Pongratz et al., 2021; Bonan, 2008; Claussen et al., 2001; Windisch et al., 2022). Increasing tree cover generally reduces surface albedo compared to grass- or cropland, leading to greater absorption of shortwave solar radiation by the surface (Bonan, 2002). Afforestation in boreal regions, thus, leads to regional biogeophysical warming due to the large albedo reduction, particularly during winter and spring when trees mask the snow-covered ground (Bonan et al., 1992; Betts, 2000). However, trees also enhance evapotranspiration through their larger leaf area and deeper root systems (Bonan, 2008), as well as through the enhancement of turbulent fluxes by their influence on surface roughness as discussed below. This is the dominant mechanism by which afforestation in tropical regions induces regional biogeophysical cooling (Claussen et al., 2001; Davin and De Noblet-Ducoudré, 2010). Additionally, trees increase surface roughness which consequently changes wind patterns and turbulent fluxes like latent and sensible heat fluxes (Burakowski et al., 2018; Sud et al., 1988). These biogeophysical effects interact with atmospheric processes and in turn impact regional to global energy and moisture fluxes, leading to impacts on cloud formation, precipitation, and surface runoff (Swann et al., 2012; Liang et al., 2022; Devaraju et al., 2015).

Previous studies have examined how forest cover changes influence local precipitation patterns, with findings that vary based on forest patch size, local climate, or specific atmospheric mechanisms (e.g., Boysen et al., 2020; Luo et al., 2022; Khanna et al., 2017; Smith et al., 2023). Large-scale deforestation has been consistently linked to reduced local precipitation in modelling studies due to reduced local moisture recycling, and vice versa for afforestation (Boysen et al., 2020; Luo et al., 2022; Spracklen et al., 2018; te Wierik et al., 2021). However, the uncertainty remains large, primarily due to model differences in moisture advection and evapotranspiration responses, alongside internal climate variability (Boysen et al., 2020; Luo et al., 2022). Observational evidence from satellite and reanalysis data corroborates these modeling results, demonstrating that historical pan-tropical deforestation (particularly at scales exceeding 50 km) has led to significant precipitation declines (Smith et al., 2023), and that the recent increase in global vegetation cover has induced precipitation increases, especially in humid regions (Zeng et al., 2018). Unlike the precipitation decline observed with large-scale deforestation, smaller deforestation patches can lead to an increase in local precipitation, for instance, due to enhanced mesoscale convection following reduced surface



roughness (Baidya Roy and Avissar, 2002; Khanna and Medvigy, 2014). These findings could imply that precipitation increase to vegetation changes at small scales may transition into precipitation decrease at larger scales (Khanna et al., 2017).

Beyond local impacts, afforestation can induce significant remote precipitation changes through modifications in atmospheric circulation and moisture transport (Swann et al., 2012; Portmann et al., 2022; Laguë et al., 2021). In particular, extratropical afforestation in the Northern Hemisphere has been shown to induce latitudinal shifts in tropical precipitation by redistributing anomalous energy, leading to a northward displacement of the Intertropical Convergence Zone (ITCZ), and vice versa for deforestation (Swann et al., 2012; Portmann et al., 2022; Liang et al., 2022; Devaraju et al., 2015). These ITCZ shifts, in turn, influence precipitation patterns across the global monsoon regions (Liang et al., 2022; Devaraju et al., 2015). However, changes in forest cover within the tropics generally exert stronger local than remote effects (Devaraju et al., 2015), similar to other forcings like aerosols (Allen et al., 2015).

Only few studies have focused on how afforestation can alter the balance between precipitation (P) and evapotranspiration (E), and consequently affect net precipitation ($P - E$), runoff (integral of $P - E$ over the drainage basin), and water availability (Cui et al., 2022; Zan et al., 2024; Lejeune et al., 2015). Net precipitation plays an important role in land-atmosphere interactions by regulating the moisture exchange between the surface and the atmosphere. Additionally, changes in $P - E$ are closely linked to atmospheric circulation changes, which drive large-scale redistribution of moisture and energy (e.g. Seager et al., 2010; Ma and Xie, 2013; Chadwick et al., 2013; Wills et al., 2016). Most previous studies about $P - E$ responses to forest cover changes have not accounted for these large-scale circulation adjustments since they relied on atmospheric moisture tracking based on the climatological circulation (Cui et al., 2022; Zan et al., 2024). As a result, their conclusions regarding afforestation-induced moisture fluxes are likely incomplete since the leading-order influence of circulation on $P - E$ is missing. The key knowledge gap, therefore, is understanding how atmospheric circulation responds to future forest cover changes and how this, in turn, shapes regional and global hydroclimate. Addressing this gap is essential given the importance of forestation as a future climate mitigation strategy.

To address this knowledge gap, we analyse hydroclimate changes in the tropics in future afforestation simulations from the Land Use Model Intercomparison Project (LUMIP; Lawrence et al., 2016). We particularly focus on local hydroclimate changes over Africa, which is projected to experience the most significant forest cover and climate changes in the LUMIP simulations but remains understudied compared to the Amazon rainforest. We aim to address the following fundamental research questions: (1) What are the biogeophysical impacts of future afforestation and avoided deforestation on the atmospheric circulation and tropical hydroclimate (P , E , $P - E$)? (2) What are the underlying physical mechanisms driving these local hydrological responses? Can these effects be attributed to changes in surface roughness, albedo, and evaporative efficiency? Answering these questions will advance our mechanistic understanding of land use and land cover change (LULCC) impacts on the hydrological cycle and will improve the understanding of the potential consequences of afforestation scenarios to reach net-zero emissions.



2 Data and methods

2.1 LUMIP simulations

We analyse changes in tropical hydroclimate to future LULCC, particularly forest cover changes, using simulations from the Land Use Model Intercomparison Project (LUMIP; Lawrence et al., 2016):

- SSP126-SSP370Lu (hereafter S1L3): This concentration-driven experiment was run with all forcings identical to the Shared Socioeconomic Pathway SSP1-2.6 (global radiative forcing of 2.6 Wm^{-2} by 2100; hereafter S1) except that the LULCC input is taken from SSP3-7.0.
- SSP370-SSP126Lu (hereafter S3L1): This concentration-driven experiment was run with all forcings identical to SSP3-7.0 (global radiative forcing of 7.0 Wm^{-2} by 2100; hereafter S3) except that the LULCC input is taken from SSP1-2.6.

These LUMIP simulations cover the period from 2015 to 2100. The differences S3L1 minus S3 as well as S1 minus S1L3 allow an analysis of the biogeophysical climate response to afforestation and avoided deforestation in a medium- and low-warming world, respectively (Table 1).

Both simulation differences are based on the projected LULCC difference from S3 to S1, which shows reduced tropical deforestation and large-scale afforestation (scenario difference in global forest area is around 1000 by 2100 (Lawrence et al., 2016)), and reduced expansion of cropland and grassland. Regions of avoided deforestation versus afforestation are shown in Fig. S1. Here, we focus on the average scenario difference averaged over the last 50 years of simulation (i.e., 2051–2100) which we refer to as AFFOREST, i.e., $\text{AFFOREST} = \frac{1}{2}(\text{S1}-\text{S1L3}) + \frac{1}{2}(\text{S3L1}-\text{S3})$. Note that AFFOREST includes both avoided deforestation and afforestation as discussed above and shown in Fig. S1, though for brevity we will sometimes refer to it as an afforestation scenario in the text. Similarly, we define the average of the simulations with S3 LULCC as BASE, i.e., $\text{BASE} = \frac{1}{2}(\text{S1L3} + \text{S3})$. Note that we focus on a 50-year timeperiod (compared to the more typical 30-year average) and the average across scenarios, which helps to increase the signal-to-noise ratio. We did not find a strong dependence of the climate response to afforestation on the low- to medium-warm background climate (see Section 3.1).

We selected seven Earth System Models (ESMs) from the models that performed these simulations based on the availability of multiple S1L3 / S3L1 ensemble members (ACCESS-ESM1-5, CESM2 and UKESM1-0-LL) or otherwise the availability of

Table 1. Summary of LUMIP experiments (i.e., difference of two simulations) analysed in this study.

Experiments	Baseline climate	Interpretation
SSP126 minus SSP126-SSP370Lu (called S1-S1L3)	SSP1-2.6	Afforestation + avoided deforestation in low warming world
SSP370-SSP126Lu minus SSP370 (called S3L1-S3)	SSP3-7.0	Afforestation + avoided deforestation in medium warming world



at least five S1 / S3 ensemble members (CNRM-ESM2-1, IPSL-CM6A-LR, MIROC-ES2L and MPI-ESM1-2-LR) to increase the signal-to-noise ratio. We had to exclude the NorESM2-LM model due to an unrealistic global mean temperature of up to 19.5°C in the S1L3 simulation and the CanESM5 model due to a different distribution of forest cover changes (including minimal forest cover change in the tropics) compared to other ESMs. We calculated the multi-model ensemble mean (MMM) from the ensemble means of these seven ESMs interpolated to a common 2.5×2.5° grid. Further details about the resolution and land modelling schemes of the chosen ESMs are given in Table 2.

Statistical significance of changes was tested by propagating the uncertainty due to internal variability through this ensemble and scenario averaging (not shown), but a simpler approach based on where 6 of 7 models agree on the sign of change gave similar answers and is used throughout the text.

Table 2. List of CMIP6 models used that performed the LUMIP simulations. The number of ensemble members, horizontal resolution and details of the land modelling scheme are given. PFT stands for plant function type. Dynamic vegetation cover in climate models refers to the simulation of vegetation changes over time in response to climate conditions, CO₂ levels, and disturbances, allowing for feedback between ecosystems and the atmosphere. Information about the horizontal resolution, number of PFTs and dynamic vegetation cover are taken from Arora et al. (2020).

Model name	Ensemble members S1L3 & S3L1	Ensemble members S1 & S3	Horizontal resolution	No. of PFTs	Dynamic vegetation cover	Irrigation	Citation LUMIP simulations
ACCESS-ESM1-5	10	10	1.875°×1.25°	13	No	No	Ziehn et al. (2021)
CESM2	3	3	0.9°×1.25°	22	No	Yes	Danabasoglu (2019)
CNRM-ESM2-1	1	5	1.4°×1.4°	16	No	No	Seferian (2019)
IPSL-CM6A-LR	1	5	2.5°×1.3°	15	No	No	Boucher et al. (2019)
MIROC-ES2L	1	5	2.81°×2.81°	13	No	No	Hajima et al. (2019)
MPI-ESM1-2-LR	1	5	1.8°×1.8°	13	Yes	No	Pongratz et al. (2019)
UKESM1-0-LL	5	5	1.875°×1.25°	13	Yes	No	Wiltshire et al. (2019)

The MMM difference in forest cover between 2051 and 2100 in AFFOREST is shown in Figure 1a. Afforestation and avoided deforestation are primarily located in central Africa, followed by North and South America, India, and central Europe. Conversely, deforestation and avoided afforestation are located in eastern China and eastern Europe. Zonal-mean changes reveal that the largest forest cover increase is located in the tropics, with a smaller increase in the boreal Northern Hemisphere (Fig. 1b). While some model differences regarding the forest cover changes exist (e.g., ACCESS-ESM1-5 shows a smaller latitudinal extent of African forest changes than other models), these differences stem mainly from variations in baseline forest cover (not shown). Decrease in crop- and grassland largely match the expansion in forest area, although the changes in grassland show notable differences across models (Fig. S2a-d). Out of the three models providing shrubland cover data, only ACCESS-ESM1-5 shows large expansions in the northern and southern boreal zones (Fig. S2e, f). Note that these simulations do contain other LULCC than forest and associated crop/grass/shrub changes such as changes in irrigation and wood harvest,



130 although only CESM2 includes the former and the latter can be related to deforestation changes. Further description of the LUMIP experiments can be found in Lawrence et al. (2016).

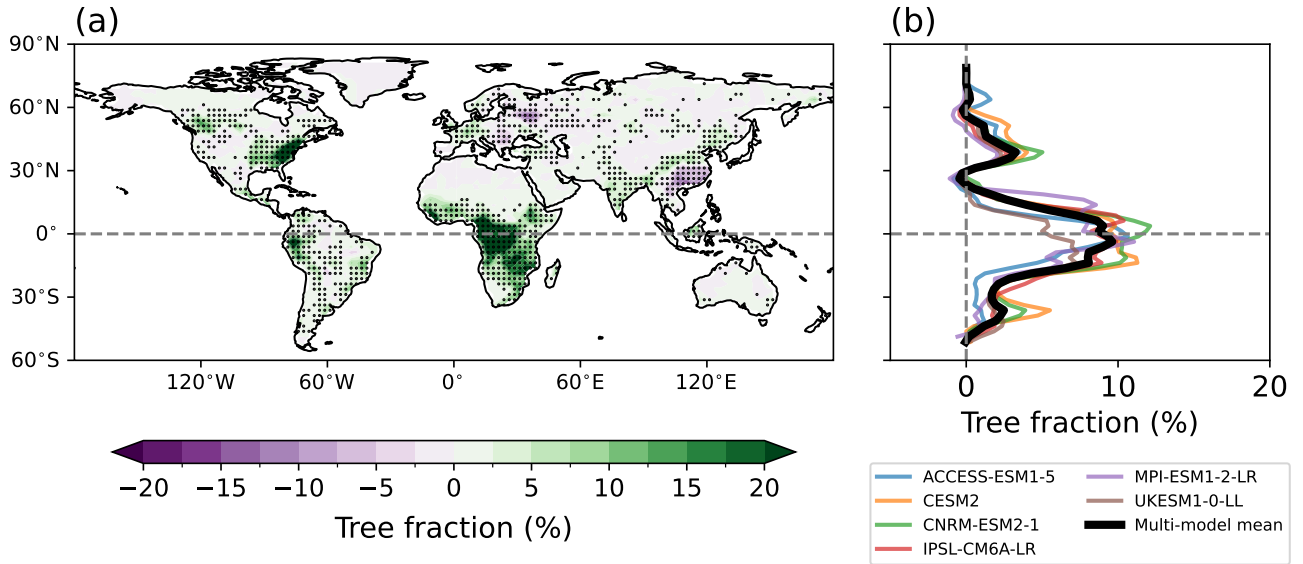


Figure 1. (a) Annual mean spatial pattern of tree fraction changes from 2051–2100 for AFFOREST. Stippling shows where 5 out of 6 ESMs agree on the sign of change. (b) Zonal-mean tree fraction change is shown for 6 ESMs and the MMM (black line). Note that MIROC-ES2L does not provide tree fraction data.

2.2 Moisture budget decomposition and related diagnostics

We apply a moisture budget decomposition, adapting the method from Seager et al. (2010) and Wills and Schneider (2016), to analyse the physical mechanism driving changes in tropical moisture fluxes and to disentangle the influence from circulation and transient eddies. Changes in the moisture budget are described by

$$\Delta(P - E) \approx \Delta MC + \Delta TE \quad (1)$$

where Δ denotes the time-mean AFFOREST difference and ΔMC and ΔTE the contributions of time-mean circulations and transient eddies to changes in moisture-flux convergence, respectively. Precipitation P and evapotranspiration E are derived from model output. The mean-circulation moisture-flux term ΔMC can be split into dynamic and thermodynamic components

140 as

$$\Delta MC = \underbrace{-\nabla \cdot \langle (\Delta \mathbf{u}) q_{Base} \rangle}_{\Delta MC_{Dyn}} - \underbrace{\nabla \cdot \langle \mathbf{u}_{Base} (\Delta q) \rangle}_{\Delta MC_{Thermo}} \quad (2)$$

where ∇ is the nabla operator on a sphere, $\mathbf{u}$ is the horizontal wind vector, q is the specific humidity and $\langle \cdot \rangle$ indicates an integration from the pressure at the top of the atmosphere to the surface normalised by the gravitational acceleration such that



the units are the same as for precipitation. The dynamic contribution ΔMC_{Dyn} arises only from changes in mean atmospheric
145 circulation and the thermodynamic contribution ΔMC_{Thermo} arises only from changes in specific humidity (Eq. 2).

The transient eddy term ΔTE in Eq. (1) captures changes in concurrent sub-seasonal variations in specific humidity and mean circulation. This term is calculated as the residual from Eq. (1), as daily data is not available for all ESMs to allow a direct calculation of this term, and it therefore also includes the small nonlinear mean-circulation term, which arises from the product of changes in humidity and circulation.

150 2.2.1 Surface and near-surface impact on moisture flux

Insights into the impact of near-surface changes on $\Delta(P - E)$ in the (sub)tropics can be gained from the omega scaling from Wills and Schneider (2015). This omega scaling is an empirical approximation of ΔMC based on the integral form of the mean value theorem and is given by

$$\Delta(P - E) \approx -\frac{1}{g} \Delta(q_s \omega_{700}) \quad (3)$$

155 with surface-specific humidity q_s and 700-hPa vertical pressure velocity ω_{700} . Note that the formula includes ω at 700 hPa, rather than 850 hPa as in Wills and Schneider (2015) and Wills and Schneider (2016), which was empirically found to better match net precipitation in our analysis and is consistent with the higher lifting condensation level (LCL) over tropical land rather than ocean (see Sec. 3.2.2.). Nonetheless, this approximation gives only a qualitative description of $\Delta(P - E)$ since it neglects the effect of transient eddies and horizontal advection, and is not defined in regions where the surface extends above
160 700 hPa.

One important influence on ω_{700} and thus $\Delta(P - E)$ is the orographic effect of moisture flow across surface pressure gradients (i.e., upward movement on the windward side and downward movement on the leeward side of orography). The surface contribution ΔS can be calculated as

$$\Delta S = -\frac{1}{g} \Delta(q_s \omega_s) = -\frac{1}{g} \Delta(q_s \mathbf{u}_s \cdot \nabla p_s) \quad (4)$$

165 where p denotes pressure, ω denotes vertical pressure velocity and the subscript s indicates values at the surface.

2.2.2 Net energy input impact on moisture flux

We also analyse changes in the net energy input (NEI) to investigate how energetic changes influence atmospheric circulation and net precipitation. In steady state, the moist static energy equation relates NEI to the divergence of moist static energy (MSE, symbol h) flux:

$$170 \text{ NEI} = \nabla \cdot \langle \mathbf{u} h \rangle. \quad (5)$$

Here, MSE is given by $h = c_p T + L_v q + \Phi$ with the specific heat capacity of air at constant pressure c_p , three-dimensional temperature T , latent heat of vaporization L_v , three-dimensional specific humidity q , and geopotential Φ . In the tropics, changes in MSE flux divergence are related to moisture flux convergence and $P - E$, because energy is transported in the opposite



direction as moisture in tropical overturning circulations, leading to anomalous $P - E$ in regions of anomalous NEI. NEI can
175 be further separated into individual components due to the surface latent heat flux (LHF), surface sensible heat flux (SHF), net
radiative energy flux into the atmospheric column (R_{NET} ; comprising both short- and long-wave contributions at the top of the
atmosphere and surface) and the atmospheric energy storage (ST):

$$NEI = LHF + SHF + R_{NET} + ST \quad (6)$$

Here, we use the convention that positive values represent fluxes into the atmospheric column. The values for LHF, SHF, R_{NET}
180 and ST can be calculated from the LUMIP data. ST is found to be negligible and is not considered in the later analysis (not
shown).

3 Results

3.1 Climate response over land

Here, we analyse the climate response over land to the future afforestation scenario (including avoided deforestation, as dis-
185 cussed in Section 2.1). In response to the afforestation, the zonal-mean evapotranspiration increases in the tropics as trees
enhance transpiration through a larger leaf area and increase the evaporative capacity due to deeper root systems (Fig. 2b; Bo-
nan, 2008). Additionally, the increase in surface roughness leads to an increase in latent heat flux and thus evapotranspiration.
The increase in evapotranspiration enhances the atmospheric moisture content, leading to an increase in precipitation over the
ITCZ region over land throughout the year, while precipitation decreases north of the equator during the extended Northern
190 Hemisphere monsoon period (i.e., March to September (MJJAS)) (Fig. 2a). The surface temperature decreases due to the large
increase in cooling by evapotranspiration, which dominates over the warming due to the land surface albedo increase in the
tropics (Fig. 2d). Notably, these changes in the hydroclimate and temperature show large agreement across models in the lati-
tudinal structure and seasonality. Additionally, these tropical climate responses are largely independent of the low- (S1-S1L3)
and medium- (S3L1-S3) warming background climate, as shown by analysing the simulations separately (Fig. S3).

195 The net precipitation ($P - E$) quantifies the exchange of moisture between the atmosphere and Earth's surface and is a
key control of surface runoff. In response to afforestation, net precipitation shows a predominant decrease north of the equator
throughout the year and south of the equator in MJJAS, accompanied by a modest increase along the ITCZ band (Fig. 2c). From
a land-climate perspective, this tropical $\Delta(P - E)$ response could be in parts seen as an increase in moisture recycling, since
it means that precipitation increases less than evapotranspiration and thus more of the moisture is sourced locally. However,
200 this response seems counterintuitive from an atmospheric dynamics perspective as NEI (given by the sum of the surface latent
heat, surface sensible heat and net radiative energy into the atmospheric column; Eq. 6) increases (Fig. 2e), which would
typically imply a strengthening of the overturning circulation and increase in net precipitation throughout the inner tropics.
The southward shift and contraction towards the equator of the ITCZ are consistent with energetic frameworks based on NEI
(Bischoff and Schneider, 2014; Byrne and Schneider, 2016), but the overall reduction of $P - E$ is not. The increase in NEI is
205 mainly determined by the increase in latent heat flux, partially compensated for by a decrease in net radiative energy into the

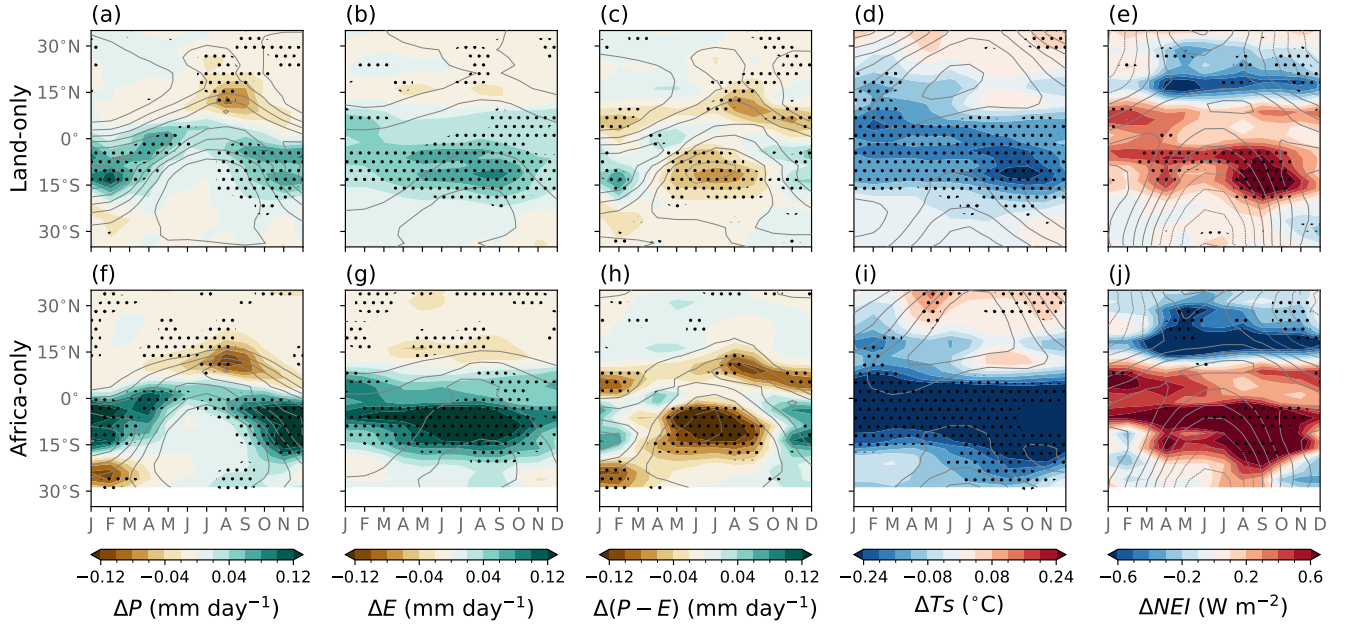


Figure 2. Zonal-mean monthly changes in (a, f) precipitation ΔP , (b, g) evapotranspiration ΔE , (c, h) net precipitation $\Delta(P - E)$, (d, i) Surface temperature ΔTs , (e, j) net energy input ΔNEI over land (upper row) and over Africa (lower row; average over land area 20°W to 50°E) in AFFOREST. Contours show the BASE values with a spacing of 1.5 mm day⁻¹ for ΔP , ΔE and $\Delta(P - E)$, a spacing of 5°C for ΔTs and of 20 W m⁻² for ΔNEI . Dashed contours indicate negative values. Stippling shows where 6 out of 7 ESMs agree on the sign of change.

atmosphere and a minimal change in sensible heat flux (Fig. S4). The net radiative changes result from changes in longwave radiation at the top of the atmosphere (ToA) and surface rather than changes in shortwave radiation (not shown). Overall, this indicates a stronger greenhouse effect over the forest, where increased water vapor and cloud cover trap outgoing radiation. As a result, the surface cools less efficiently (Fig. S4f) and less longwave radiation escapes at the top of the atmosphere (Fig. S4e).

210 The opposing effects in ToA and surface longwave radiation combine with latent heat flux to give the net effect in NEI.

This study thus seeks to identify the mechanisms driving the $\Delta(P - E)$ pattern over the tropics from a dynamics perspective and to reconcile the apparent mismatch between the tropical $\Delta(P - E)$ and ΔNEI . Later analysis focuses on the response over Africa which we identify to be the dominant region influencing the hydroclimate, temperature, and NEI response over the tropics (Fig. 2f-j, S5). This finding aligns with projected differences in forest cover between S1 and S3, which are most pronounced over Africa (Fig. 1a).

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3.2 Mechanistic insights into moisture flux changes over Africa

3.2.1 Moisture budget analysis

The following analysis focuses on hydroclimate changes over Africa and possible underlying mechanisms. The analysis is split up into the two extended monsoonal seasons influencing the monsoon precipitation over western Africa (May to September, 220 MJJAS) and southern Africa (November to March, NDJFM) (Douville et al., 2021). The spatial pattern of precipitation change shows an increase in equatorial Africa in MJJAS, while precipitation decreases over large parts of the western African monsoon region (Fig. 3a; cf. climatology in Fig. S6a). During NDJFM, precipitation increases over the southern African monsoon region (Fig. 4a; cf. climatology in Fig. S6e). Evapotranspiration increases in both seasons over central Africa, the region of largest afforestation and avoided deforestation between the SSPs (Fig. 3b, 4b). This leads to a tripole dry-wet-dry pattern in net 225 precipitation over Africa in MJJAS and NDJFM, with varying extents of the wetting (Fig. 3c, 4c). Thus, the climatological dry ($P - E$) region over southern Africa and wet region over central Africa are strengthened in MJJAS (Fig. 3c; cf. climatology in Fig. S6c), as are the climatological wet ($P - E$) region over southern Africa and dry region over western Africa in NDJFM (Fig. 4c; cf. climatology in Fig. S6g).

We perform a moisture budget decomposition as described in Section 2.2 to investigate the underlying dynamical and ther- 230 modynamical changes, and associated mechanism leading to the $\Delta(P - E)$ pattern. The net precipitation changes in both monsoonal seasons are primarily driven by changes in the mean circulation moisture flux ΔMC (Fig. 3d, 4d). There is relatively large agreement among the models regarding the magnitude and spatial pattern of the ΔMC term, albeit with some spatial variability across individual models (Fig. S7, S8). Nonetheless, in all models, the ΔMC term emerges as the dominant contributor to the $\Delta(P - E)$ response during both seasons (Fig. S7, S8). In particular, the dynamic component contributes 235 to this ΔMC pattern (Fig. S9b,e), while the thermodynamic component is small despite large and robust changes in specific humidity in central Africa (Fig. S9c, f; cf. specific humidity changes in Fig. S10a, d). Transient eddy changes are also smaller than dynamic ΔMC changes, only significantly contributing to the net precipitation drying in southeastern Africa in MJJAS (Fig. 3e) and wetting over central Africa in NDJFM (Fig. 4e). Thus, the mechanism(s) explaining the net precipitation changes over Africa are of dynamic origin (i.e., related to changes in the time-mean circulation), rather than related to 240 transient-eddy or thermodynamic changes.

3.2.2 Surface and lower-tropospheric changes

Insights into the impact of near-surface changes and the atmospheric levels relevant to explaining the ΔMC and thus $\Delta(P - E)$ pattern can be gained from the empirical omega scaling (Eq. 3). The omega scaling shows overall agreement with the net precipitation changes (Fig. 3f, 4f). Note that the formula includes ω at 700 hPa, in contrast to 850 hPa as in Wills and Schneider 245 (2015) and Wills and Schneider (2016), which was empirically found to better match net precipitation in our analysis and is consistent with the higher LCL over tropical land rather than ocean (Fig. S11). Additionally, while the formula for the omega scaling incorporates specific humidity changes (Eq. 3), we found the contribution of the thermodynamic effects to $\Delta(P - E)$ to be minimal. This emphasizes the dominant impact of dynamical changes in convergence below 700 hPa on $\Delta(P - E)$.

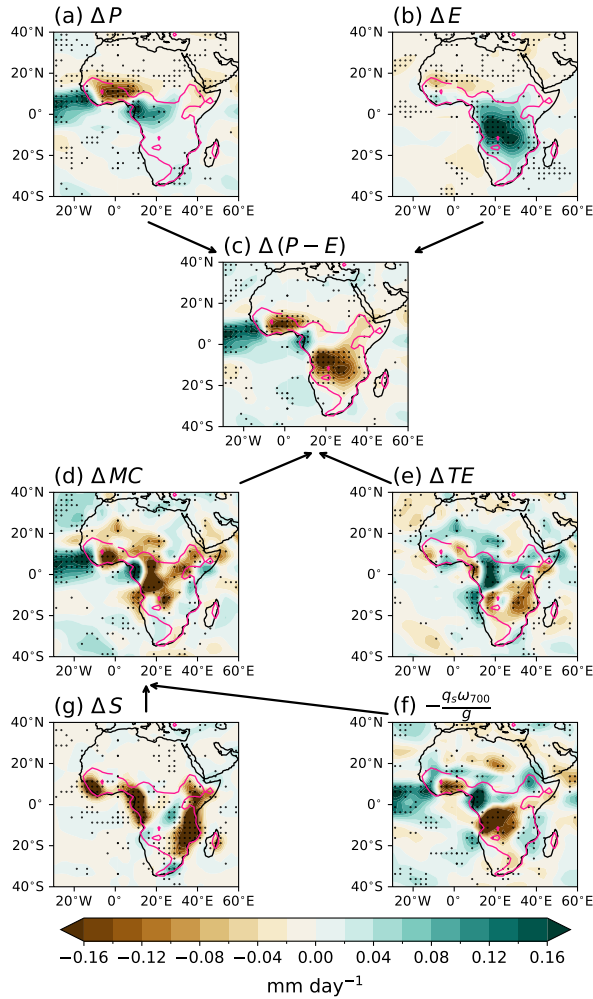


Figure 3. Spatial maps of MJJAS changes in (a) precipitation ΔP , (b) evapotranspiration ΔE , (c) net precipitation $\Delta(P - E)$, (d) mean circulation moisture flux term ΔMC , (e) transient eddy term ΔTE , (f) omega scaling and (g) surface term ΔS over Africa in AFFOREST. Pink contours enclose the regions where there are at least 5% of afforestation. Stippling shows where 6 out of 7 ESMs agree on the sign of change. Arrows show the relationship between the different variables.

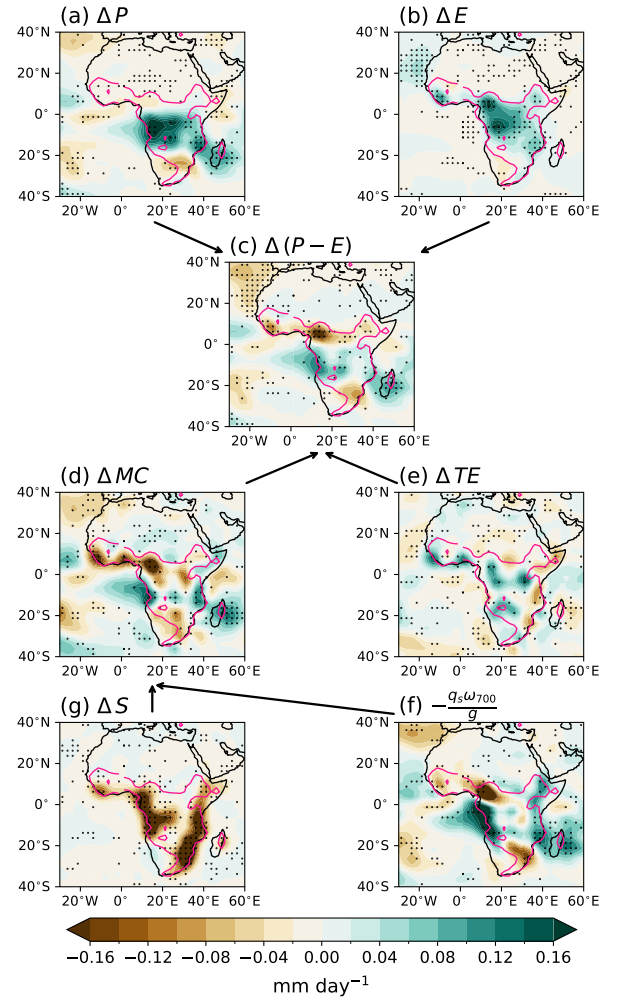


Figure 4. Spatial maps of NDJFM changes in (a) precipitation ΔP , (b) evapotranspiration ΔE , (c) net precipitation $\Delta(P - E)$, (d) mean circulation moisture flux term ΔMC , (e) transient eddy term ΔTE , (f) omega scaling and (g) surface term ΔS over Africa in AFFOREST. Pink contours enclose the regions where there are at least 5% of afforestation. Stippling shows where 6 out of 7 ESMs agree on the sign of change. Arrows show the relationship between the different variables.

An important contributor to ω_{700} and thus $\Delta(P - E)$ is the orographic effect of moisture flow across surface pressure gradients, which we quantify with the surface term ΔS (Eq. 4). Following afforestation, ΔS shows a consistent decrease near the mountain ridges along the southeast, southwest, and west coasts of Africa during both seasons (Fig. 3g, 4g; cf. climatology



in Fig. S6d, h). This response can be attributed to an afforestation-induced increase in surface drag, which slows down moisture-laden surface winds originating from the ocean (Fig. 5a, c). The wind speed reduction is around 8% averaged across Africa (Fig. 5b, d), although it reaches above 15% in central Africa (Fig. S12a, c). This is accompanied by a concurrent increase in
255 wind stress of a similar magnitude (Fig. S12b, d), indicating that the wind speed reduction results from an increase in roughness length due to afforestation.

The reduced wind speed hinders moisture transport and diminishes upward motion along the topographic slopes by reducing the surface vertical wind component $\omega_s \equiv \mathbf{u}_s \cdot \nabla p_s$ (shown by the pink contours in Fig. 5a, c). This contributes to a reduction in vertical motion at the LCL, which reduces topographically induced precipitation and (together with the increase in evapotranspiration) results in a negative signal in ΔS . Note that this signal is not directly visible in the precipitation pattern (Fig. 3a, 4a) due to the modification of vertical velocities between the surface and the top of the LCL, as described in the following section. This surface effect extends throughout the lower troposphere, with wind speed reductions over Africa falling off approximately logarithmically until around 700 hPa (Fig. 5b, d). Note that we interpolated the vertical velocity profiles to sigma coordinates (i.e., pressure divided by the grid-cell surface pressure in BASE) to show the effect on surface values more clearly, since in
265 pressure coordinates the pressure levels correspond to different heights above the surface. Sigma coordinates are then displayed in units of pressure by multiplying by the area-averaged surface pressure in BASE. Overall, while the surface-drag induced slow-down of lower-tropospheric winds and associated vertical velocity have a substantial influence on $\Delta(P - E)$, there are likely competing effects contributing to the observed discrepancies between ΔS and $\Delta(P - E)$.

3.2.3 Free-tropospheric changes

270 We next investigate free-tropospheric mechanisms influencing $\Delta(P - E)$ as well as the reasons for the apparent discrepancy with NEI changes. For this, we examine spatial patterns of NEI over Africa and vertical velocity profiles over the alternating dry-wet-dry regions of $\Delta(P - E)$ in AFFOREST (Fig. 6, 7).

Over western and southern Africa, $P - E$ decreases in MJJAS and NDJFM (Fig. 6a, 7a). The afforestation-induced changes in vertical velocity (light blue lines in Fig. 6c,e; 7c,e) oppose the climatological motion (dark blue line in Fig. 6c,e; 7c,e). This
275 leads to a weakening of the upwards motion over these regions and thus a negative $\Delta(P - E)$ response. The western Africa response in NDJFM and the southern Africa response in MJJAS show vertical velocity anomalies concentrated in the lower troposphere, indicating that the surface drag-induced change is the dominant effect impacting the vertical velocity profile over western and southern Africa in these seasons. The MJJAS response over western Africa shows small changes in the lower-troposphere, albeit with large spatial differences within the region (Fig. 3f), indicating the $P - E$ response in this region results
280 from a complex interplay of different mechanisms, e.g., including the influence of a strengthened Saharan heat low (Fig. S10b, c).

Conversely, over central Africa, $P - E$ increases in MJJAS and NDJFM (Fig. 6a, 7a). This response is related to the strong increase in NEI over central Africa, which is dominated by an increase in latent heat flux and the associated longwave radiative effects in a moister and cloudier atmosphere, during MJJAS and NDJFM (Fig. 6b, 7b, S4). This excess NEI strengthens the
285 climatological upward motion in the free troposphere (i.e., AFFOREST changes in the direction of BASE values in Fig. 6d

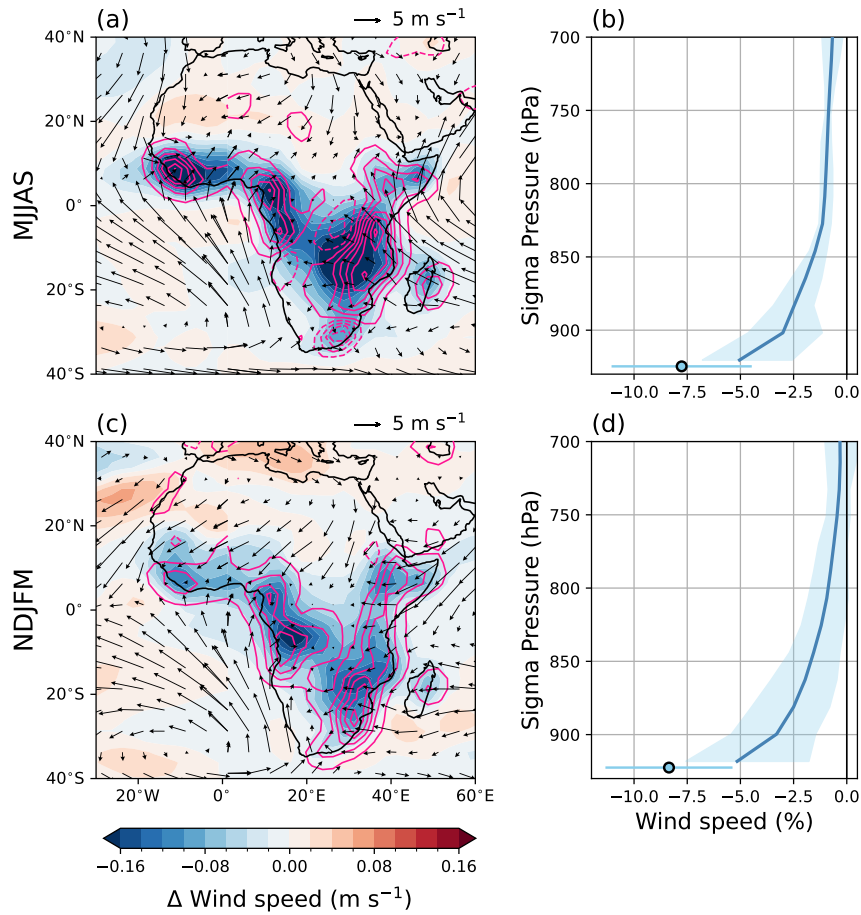


Figure 5. (a, c) Spatial maps of changes in surface wind speed (filled contours) over Africa for AFFOREST in MJJAS (upper row) and NDJFM (lower row). Pink contours show changes in surface vertical velocity ω_s with solid (dashed) contours indicating positive (negative) values. Vectors show BASE surface winds. (b, d) Vertical profile of percentage changes in wind speed over Africa (land region of 37°S-15°N, 20°W-50°E) interpolated to sigma coordinates. The sigma coordinates are calculated as pressure divided by the grid-cell surface pressure in BASE and multiplied by the area-averaged surface pressure in BASE to get sigma coordinates in pressure units. The envelope shows the 25th to 75th percentile model spread. The light blue dot shows the percentage changes in surface wind speed (positioned at the average surface pressure in BASE) with the error bar showing the 25th-75th percentile model spread.

and 7d) by enhancing convection. Consequently, the climatological wet $P - E$ region over central Africa is strengthened (Fig. 6d, 7d; cf. climatology in S6c, d). The vertical velocity and $\Delta(P - E)$ response to excess NEI over central Africa also agrees with previous findings from Byrne and Schneider (2016) that an increase of NEI in the ITCZ region increases the vertical velocity and leads to a narrowing of the ITCZ (also seen in Fig. 2c, h). During NDJFM, the increase in NEI is overall smaller than in MJJAS but the NEI maxima over Africa are located closer to the deep convective region (i.e., region of largest



climatological NEI), leading to a larger region of $\Delta(P - E)$ wettening. Note that near-surface vertical velocity is weakened over central Africa due to the surface drag-induced effect described previously.

These findings reveal that net precipitation changes over Africa are driven by the complex interplay of afforestation effects on lower tropospheric (until around 700 hPa) circulation changes, through changes in surface drag, and on deep convection, through changes in NEI. While increased surface drag reduces $P - E$ in orographic regions by slowing down surface and lower-tropospheric winds and associated surface vertical velocities, the anomalous NEI modulates $P - E$ by strengthening convective upper-tropospheric circulations. The regional sign of $\Delta(P - E)$ then depends on the dominance of the surface drag or energetic effect over different African subregions. These findings help to reconcile the NEI pattern with the $\Delta(P - E)$ response. Another way of thinking about how our results reconcile the NEI and $P - E$ change is to consider that the idea that positive NEI anomalies lead to positive net precipitation anomalies assumes vertical velocity changes to be the same sign throughout the atmospheric column, which is not true in AFFOREST due to differing drivers of change in the upper and lower troposphere. The changes in the vertical profile of vertical velocity correspond to a change in the gross moist stability, which determines how deep-convective circulations respond to NEI.

3.2.4 Summary of mechanisms governing net precipitation response

Here, we summarise the surface drag and energetic mechanisms we determined to influence moisture fluxes and $P - E$ following afforestation (Figure 8). In the base state without trees, evapotranspiration is relatively low over crop- and grassland (Figure 8a). Orographic precipitation occurs as moisture-laden winds ascend topographic slopes, reach the LCL and condense into clouds. Convective precipitation, on the other hand, is closely linked to surface evapotranspiration, which increases the propensity for convection and associated precipitation. The resulting net precipitation is positive in most regions, leading to runoff and water availability. Following afforestation, two competing mechanisms shape the hydroclimate response over Africa. The first is the surface drag effect, in which increased surface drag following afforestation slows moist onshore winds, in turn reducing near-surface and lower-tropospheric vertical velocity (Figure 8b). This leads to a suppression of orographic precipitation in response to afforestation, such that precipitation increases less than evapotranspiration, leading to a decrease in net precipitation compared to the base state. The second mechanism is the energetic effect (Figure 8c), where afforestation increases the net energy input to the atmosphere due to greater latent heat flux (NEI_{LHF}), partially offset by increased radiative cooling of the atmosphere (NEI_{Rnet}). This leads to a strengthening of the climatological vertical upward motion in the free troposphere. The associated enhancement of convective precipitation, together with increased tree-driven evapotranspiration, results in an increase in net precipitation compared to the base state. Our results indicate that net precipitation changes over Africa in response to afforestation are driven by a complex interplay between lower- and upper-tropospheric circulation changes. Both the energetic and the surface-drag effect are active over the entire African domain to afforestation, however the resulting net precipitation response depends on which effect is dominant at around 700hPa over a certain region. The surface drag effect is the dominant driver of the net precipitation decrease over western and southern Africa during MJJAS and NDJFM, although modulated by the energetic effect. Conversely, the energetic effect is the dominant driver of the net precipitation increase over central Africa, though the surface drag effect still influences near-surface vertical velocity. Overall, this

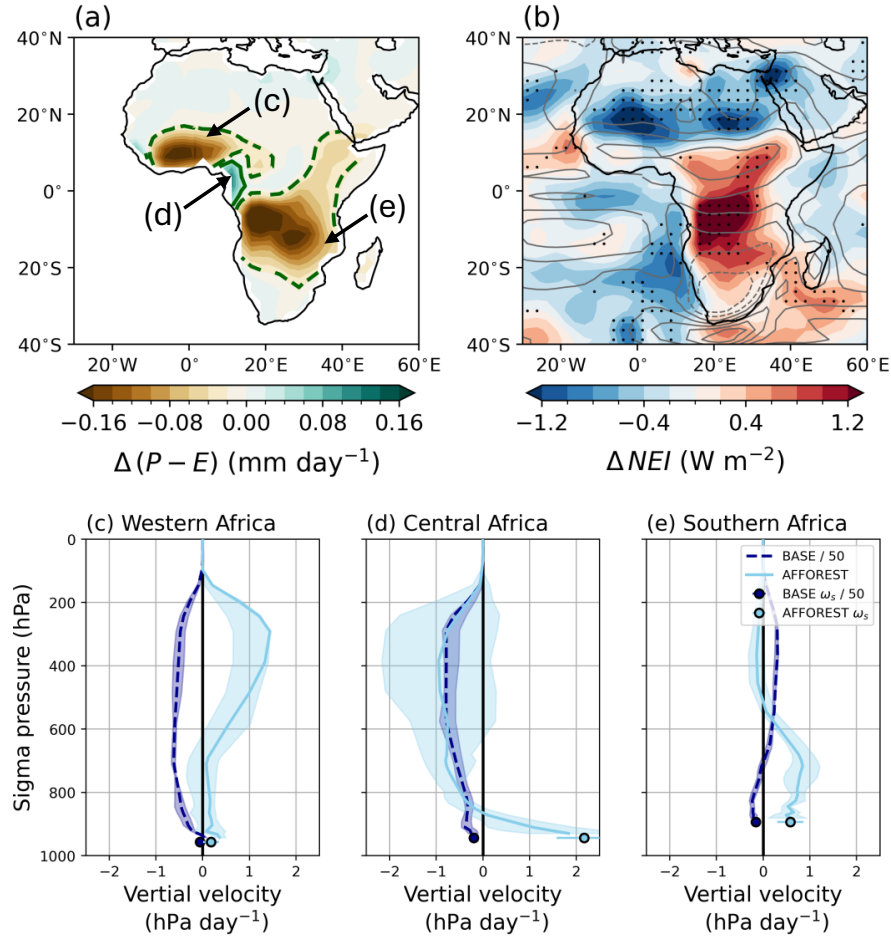


Figure 6. Spatial maps of MJJAS changes in (a) net precipitation $\Delta(P - E)$ and (b) net energy input ΔNEI over Africa in AFFOREST. In (b), contours show BASE values of NEI with a spacing of 20 W m⁻². (c-e) Vertical profile of BASE values (dark blue) and changes following AFFOREST (light blue) in vertical velocity ω averaged over the western, central and southern African regions indicated in (a). The dark blue dot shows the BASE value and light blue one the AFFOREST value of vertical velocity at the surface ω_s with error bars showing the 25-75th percentile. Note that BASE values have been divided by 50 to display them on the same axes and the profiles are interpolated to sigma coordinates. The sigma coordinates are calculated as pressure divided by the grid-cell surface pressure in BASE and multiplied by the area-averaged surface pressure in BASE to get sigma coordinates in pressure units. The surface values (dots) are positioned at the average surface pressure in BASE over the relevant region.

dynamic interaction between local surface processes and large-scale atmospheric energy balances explains the spatial patterns of hydroclimate change seen in our study.

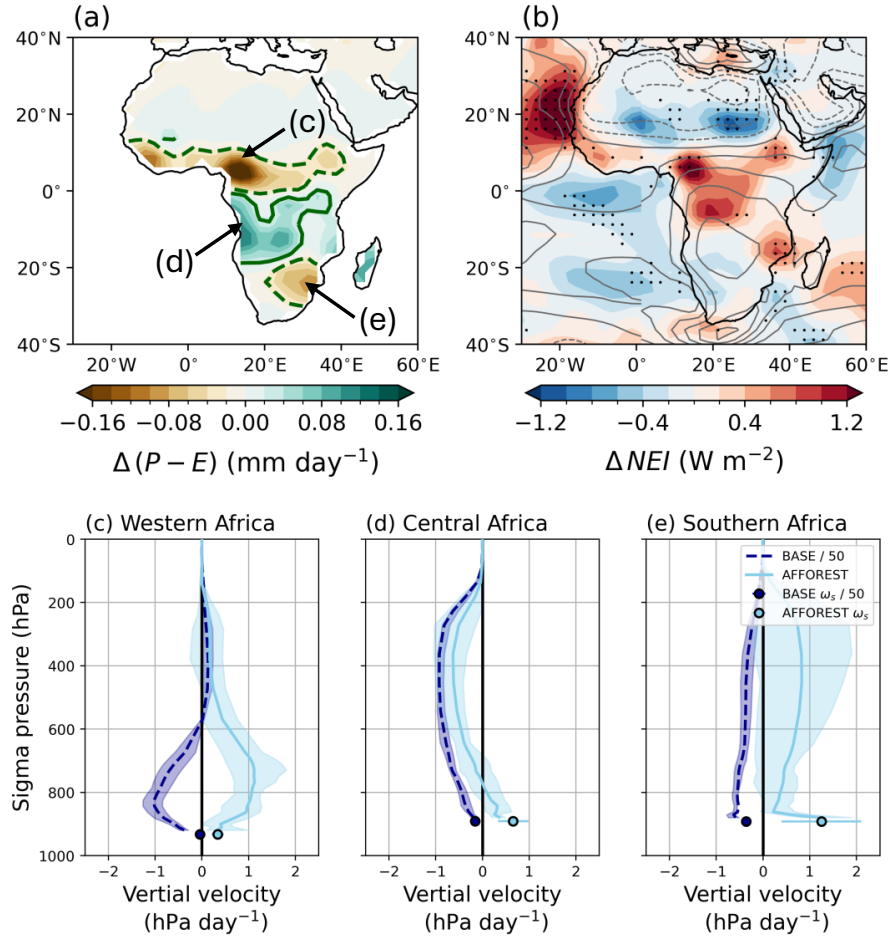


Figure 7. Spatial maps of NDJFM changes in (a) net precipitation $\Delta(P - E)$ and (b) net energy input ΔNEI over Africa in AFFOREST. In (b), contours show BASE values of NEI with a spacing of 20 W m⁻². (c-e) Vertical profile of BASE values (dark blue) and changes following AFFOREST (light blue) in vertical velocity ω averaged over the western, central and southern African regions indicated in (a). The dark blue dot shows the BASE value and light blue one the AFFOREST value of vertical velocity at the surface ω_s with error bars showing the 25-75th percentile. Note that BASE values have been divided by 50 to display them on the same axes and the profiles are interpolated to sigma coordinates. The sigma coordinates are calculated as pressure divided by the grid-cell surface pressure in BASE and multiplied by the area-averaged surface pressure in BASE to get sigma coordinates in pressure units. The surface values (dots) are positioned at the average surface pressure in BASE over the relevant region. Also note that the regions marked as western, central and southern Africa are not the same as in Fig. 6.

4 Discussion and conclusions

In this study, we analysed hydroclimate changes in the tropics under future afforestation scenarios using simulations from the Land Use Model Intercomparison Project (LUMIP; Lawrence et al., 2016). Our primary focus is on hydroclimate and

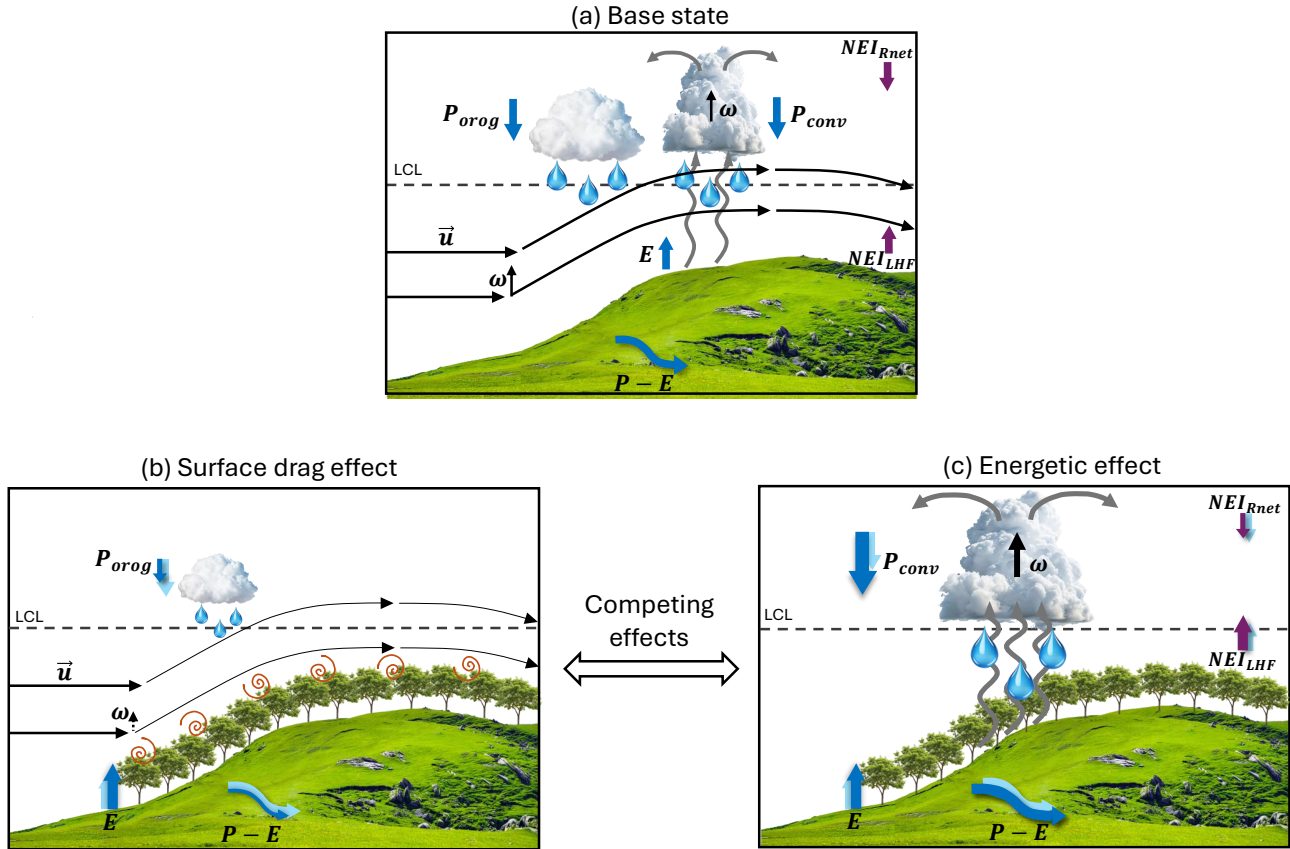


Figure 8. Schematic illustrating (a) the mechanisms affecting the hydroclimate in the base state without trees, along with the impacts of (b) surface drag and (c) energetic changes on moisture flux resulting from afforestation. Moisture fluxes (P , E , $P - E$) are represented by dark blue arrows, pointing in the direction of flow. In (b) and (c) the climatological moisture and energetic fluxes from (a) are depicted as light blue arrows for reference. The wind flow over topography is shown as black arrows (labelled $\vec{u}$) and vertical winds are depicted by upward-pointing black arrows (labelled ω). Convective fluxes are represented by curvy grey arrows, with outflow indicated by curved grey arrows at the top of the cloud. Stronger fluxes and circulation are represented by thicker and longer arrows, while weaker fluxes are indicated by thinner and shorter arrows. Similarly, stronger precipitation intensity is shown by larger raindrop and cloud symbols and vice versa for lower precipitation intensity. The lifting condensation level is labelled as LCL, orographic precipitation as P_{orog} and convective precipitation as P_{conv} . NEI_{LHF} and NEI_{Rnet} refer to the latent heat flux and the net radiative energy component of the net energy input NEI, respectively.

330 atmospheric changes over Africa, which is simulated to undergo significant hydrological changes in these simulations but remains understudied compared to the Amazon rainforest. We identify three key findings:

1. Afforestation and avoided deforestation lead to a robust increase in precipitation and evapotranspiration, alongside widespread decreases in net precipitation (precipitation minus evaporation) in the tropics, particularly over Africa.



2. The tropical hydroclimate response to future forest cover change is largely independent of the background climate under low- and medium-warming scenarios.
3. The changes in net precipitation over Africa are driven by the competing effects of surface drag-induced reduction of lower-tropospheric winds and net energy input-induced strengthening of deep-convective upper-tropospheric circulations.

Here, we compare these findings to existing literature to contextualize our results.

Our results show that the hydroclimate response is most pronounced in the regions of the greatest forest cover changes, namely over Africa (Fig. 1, 2). The African climate response dominates the overall tropical hydroclimate response over land, highlighting the larger regional versus remote responses to tropical forest cover changes, although net precipitation anomalies provided evidence of moisture transport by the large-scale circulation. This aligns with the findings of Devaraju et al. (2015) who demonstrated that while large-scale deforestation in the northern mid- and high latitudes can influence the ITCZ remotely, tropical forest cover changes primarily induce regional hydroclimate responses. Across models, we find a robust increase in evapotranspiration following afforestation, which can be attributed to enhanced leaf area and deeper root systems (Bonan, 2008), as well as larger turbulent fluxes due to surface roughness changes (Burakowski et al., 2018; Sud et al., 1988). Our results further show a consistent increase in precipitation in response to afforestation which is in agreement with satellite-based studies reporting significant local increases in precipitation to recent increases in global vegetation cover (Zeng et al., 2018) and vice versa for tropical deforestation (Smith et al., 2023). These findings are further supported by previous single model modelling studies demonstrating that deforestation in Africa leads to a local reduction in precipitation (e.g. Varejao-Silva et al., 1998; Semazzi and Song, 2001; Nogherotto et al., 2013). However, multi-model studies for instance investigating the precipitation response to large-scale idealised deforestation find that models do not show consistent precipitation changes in many regions (Luo et al., 2022; Boysen et al., 2020). Thus, the consistent precipitation (and net precipitation) changes over Africa across multiple models in response to future forest cover changes represents a significant new insight. Taken together, these findings emphasize the critical role of regional forest cover in modulating regional hydroclimate conditions.

We find that the temperature and hydroclimate responses to afforestation in the tropics are largely independent of the background climate under low- to medium-warming scenarios (Fig. S3). Previous studies have highlighted that the biogeophysical cooling effect of vegetation is more significant in a warming world (Pitman et al., 2011; Hua and Chen, 2013), although these studies primarily concentrated on the snow-albedo feedback of LULCC in mid- and high-latitudes which weakens under increased CO₂ emissions. Studies incorporating both biogeophysical and biogeochemical effects suggest that the cooling effect of an increase in global vegetation cover is enhanced under high-emission scenarios due to the CO₂ fertilisation effect (Alkama et al., 2022; Kim et al., 2017), with around 20% of the total cooling attributable to biogeophysical processes (Alkama et al., 2022). However, when compared to overall global warming in a fractional sense, the cooling potential of LULCC declines in high-emission scenarios (Alkama et al., 2022; Loughran et al., 2023). Despite these insights, few studies have specifically addressed how hydroclimate responses to LULCC depend on the background climate, particularly in the tropics. To the best



of our knowledge, this is the first study that documents the mean-state independence of the afforestation-induced hydroclimate response under low-to-medium climate warming.

The mechanisms underlying precipitation changes due to forest cover alterations have been explored by some previous studies (Khanna and Medvigy, 2014; Crook et al., 2023; Nogherotto et al., 2013; Abiodun et al., 2008; Lejeune et al., 2015; Sud et al., 1988, 1996; Cheng and Mccoll, 2023; Bell et al., 2015; Luo et al., 2022; Arnault et al., 2023; Akkermans et al., 2014). Research on Amazonian deforestation suggests that small-scale or patchy deforestation increases sensible heat flux while reducing latent heat flux, resulting in surface warming and mesoscale convection that can enhance local precipitation (Khanna and Medvigy, 2014; Crook et al., 2023; Nogherotto et al., 2013; Akkermans et al., 2014). A recent study by Qin et al. (2025) found that different mechanisms impact rainfall during the dry and wet season after Amazonian deforestation. During the wet season, they found that deforestation warms the land surface, creating a low pressure area that draws in moisture and increases precipitation over deforested patches. During the dry season, they found that deforestation reduces evapotranspiration and lower atmospheric moisture levels leading to a decrease in precipitation across a wider region. However, these previous studies largely lack a more quantitative analysis of the underlying mechanisms through atmospheric dynamics methods. Additionally, these studies all consider precipitation, which is directly influenced by the evapotranspiration changes associated with forest cover changes, and our results indicate that net precipitation changes (associated with moisture flux changes) often change in the opposite sense as precipitation, indicating the need to consider net precipitation in studies of the hydroclimate response to forest cover change.

The impact of forest cover changes on net precipitation has been investigated by only a few studies, including some using atmospheric moisture tracking models (Zan et al., 2024; Cui et al., 2022) and some using regional (Lejeune et al., 2015) or global climate models (Luo et al., 2022). Previous studies using moisture tracking models, which rely on climatological circulation patterns, do not account for large-scale circulation adjustments. However, our findings indicate that neglecting these circulation adjustments leads to an incomplete understanding of the hydroclimatic impacts of forest cover changes, as circulation responses play a crucial role in determining net precipitation variations (Fig. 3, 4, S9). Similarly, Luo et al. (2022) found that net precipitation changes over Africa to an idealised deforestation scenario are primarily influenced by mean circulation moisture flux based on a moisture budget analysis. While their findings agree with our results, they find less model agreement, which we speculate is due to the larger relative importance of extratropical afforestation in their simulations.

Our results shows that the surface roughness increase due to afforestation reduces net precipitation, which aligns with previous work focusing on roughness-induced changes in precipitation (Bell et al., 2015; Crook et al., 2023; Lejeune et al., 2015; Sud et al., 1988, 1996). These studies found that the reduction in surface roughness due to deforestation enhances lower-tropospheric wind speeds, which can enhance moisture transport from the ocean to deforested regions (Bell et al., 2015; Crook et al., 2023; Lejeune et al., 2015; Sud et al., 1988, 1996). For example, Bell et al. (2015), determined that complete deforestation of the Congo primarily influences regional precipitation via changes in surface roughness, which strengthens horizontal winds over the deforested area. The intensified winds enhanced moisture transport from the Atlantic Ocean and western Congo, leading to increased moisture convergence west of the Rift Valley highlands and consequently greater rainfall over the eastern Congo Basin. Our results support the decrease in surface wind speed due to afforestation (Fig. 5, S12). However, in contrast to



previous studies we further link these surface drag-induced circulation changes to their impact on orographic precipitation and subsequent net precipitation changes (Fig. 3f, 4f), which has not been previously documented.

Our study demonstrates the competition between the surface drag effect found in previous studies (Bell et al., 2015; Crook
405 et al., 2023; Lejeune et al., 2015; Sud et al., 1988, 1996) and the energetic effect, due to change in net energy input to the atmosphere, in setting net precipitation changes over Africa. In particular, we identify a novel climate dynamical mechanism where $P - E$ changes occur opposite to both precipitation (P) and NEI in certain regions (i.e., southern and western Africa) due to the competition between the surface drag and energetic effects, while the NEI-driven changes dominate over climatologically ascending regions (i.e., central Africa) (Fig. 6, 7). Only one study by Lejeune et al. (2015) focusing on Amazonian deforestation
410 tion has explicitly examined the role of convective processes alongside the influence of surface roughness on net precipitation response. They found a longitudinal precipitation dipole and increase in net precipitation over Amazonia to regional deforestation. They attributed the precipitation response to a combination of strengthened moisture-carrying winds due to decreased surface roughness, and reduced vertical velocity and convection due to reduced net radiation at the surface and thus reduced latent heat flux. However, they acknowledge that the use of a regional climate model might dampen large-scale circulation
415 changes and related precipitation changes, due to the imposed lateral boundary conditions. In our study focused on Africa, we similarly find that surface roughness changes and convective changes are important (Fig. 8). Furthermore, we extend their previous work by applying robust methods like the moisture budget decomposition, linking these convective changes to large-scale energy budget modifications, and showing that a similar mechanism operates across tropical Africa. Overall, our findings underscore the necessity of considering both surface roughness and energetic mechanisms when evaluating the hydrological
420 impacts of land cover changes.

Here, we present the first comprehensive analysis of the tropical hydroclimate impacts and underlying mechanisms associated with future "realistic" afforestation scenarios over Africa. Additionally, we were able to reduce the model and internal variability uncertainty in the results by using multiple ESMs and multiple ensemble members for each model, in contrast to previous studies. While some previous research has highlighted the importance of surface roughness in influencing precipitation,
425 we identify the competing effects of surface roughness and energetic effects on net precipitation. However, some uncertainties remain. Firstly, while we attribute the net precipitation changes to regional variations in forest cover over Africa, alterations in forest cover across other regions may induce non-local effects that modulate (although certainly not dominate) the hydroclimate response to regional forest cover over Africa. Secondly, there might be interactions between the identified surface drag and the energetic effects, since near-surface convergence, which we relate to surface drag changes, can also influence convection and
430 so the energetic effect. Thirdly, differences remain in the mean circulation change patterns across models, although the primary drivers of net precipitation change remain the same across models. Fourthly, we were unable to perform single-perturbed parameter simulations, such as altering surface drag alone, to further validate our mechanistic analysis. However, it is important to note that most modeling approaches modify surface roughness by adjusting vegetation height (e.g. Bell et al., 2015; Lague et al., 2019), which simultaneously affects momentum, heat, and moisture fluxes (i.e., including evaporation). This interdependence
435 makes it challenging to isolate the effects of surface drag alone. Finally, we focused solely on biogeophysical effects and



there may be competing biogeochemical factors that influence the hydroclimate, which would require additional simulations incorporating both biogeophysical and biogeochemical effects.

Overall, our study contributes to an improved understanding of land-atmosphere interactions and associated hydroclimate changes in the tropics, particularly Africa. These findings underscore the importance of forest-circulation-climate feedbacks for accurately predicting hydroclimate responses to afforestation, with significant implications for sustainable forest and water resource management. Additionally, our finding that afforestation reduces net precipitation implies (with opposite sign) an increase in net precipitation and the potential for flooding in response to deforestation, which is crucial for climate change adaptation management. Given that most current afforestation strategies tend to overlook or underestimate the potential hydroclimate effects, future scientific advances are essential to better assess the atmospheric and hydrological impacts of large-scale forest cover changes. This will ultimately help guide more effective afforestation strategies aimed at achieving net-zero emissions.

Data availability. The LUMIP simulations used in this study are published by the CMIP6 archive available via the Earth System Grid Federation (ESGF) at <https://esgf-node.llnl.gov/projects/esgf-llnl/>. The schematic was created in power point using vector images from Vecteezy.

Author contributions. All authors contributed to the conceptualization of this study. NLSF performed the data analysis and calculations, visualised the results, and wrote the initial draft. RCJW and SJDH contributed to the interpretation and discussion of the results, and edited the manuscript.

Competing interests. The contact author has declared that none of the authors has any competing interests.

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