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REVIEW

Advances in understanding large-scale responses of the water cycle to climate change

Richard P. Allan,¹  Mathew Barlow,²  Michael P. Byrne,^{3,4}  Annalisa Cherchi,⁵ 
 Hervé Douville,⁶  Hayley J. Fowler,⁷  Thian Y. Gan,⁸  Angeline G. Pendergrass,⁹ 
 Daniel Rosenfeld,^{10,11}  Abigail L. S. Swann,¹²  Laura J. Wilcox,¹³  and Olga Zolina^{14,15}

¹Department of Meteorology and National Centre for Earth Observation, University of Reading, Reading, United Kingdom.

²Department of Environmental Earth and Atmospheric Sciences, University of Massachusetts Lowell, Lowell, Massachusetts.

³School of Earth and Environmental Science, University of St Andrews, St Andrews, United Kingdom. ⁴Department of Physics, University of Oxford, Oxford, United Kingdom. ⁵Istituto Nazionale di Geofisica e Vulcanologia Sezione di Bologna, INGV, Bologna, Italy. ⁶Centre National de Recherches Météorologiques, Météo-France/CNRS, Toulouse, France. ⁷University of Newcastle, Newcastle upon Tyne, United Kingdom. ⁸University of Alberta, Edmonton, Alberta, Canada. ⁹National Center for Atmospheric Research, Boulder, Colorado. ¹⁰Institute of Earth Sciences, The Hebrew University of Jerusalem, Jerusalem, Israel. ¹¹School of Atmospheric Sciences, Nanjing University, Nanjing, China. ¹²University of Washington, Seattle, Washington. ¹³National Centre for Atmospheric Science, Department of Meteorology, University of Reading, Reading, United Kingdom. ¹⁴L'Institut des Géosciences de l'Environnement/Centre National de la Recherche Scientifique, L'Université Grenoble Alpes, Grenoble, France. ¹⁵P. P. Shirshov Institute of Oceanology, Russian Academy of Sciences, Moscow, Russia

Address for correspondence: Richard P. Allan, Department of Meteorology and National Centre for Earth Observation, University of Reading, Whiteknights, Earley Gate, PO Box 243, Reading RG6 6BB, United Kingdom. r.p.allan@reading.ac.uk

Globally, thermodynamics explains an increase in atmospheric water vapor with warming of around 7%/°C near to the surface. In contrast, global precipitation and evaporation are constrained by the Earth's energy balance to increase at ~2–3%/°C. However, this rate of increase is suppressed by rapid atmospheric adjustments in response to greenhouse gases and absorbing aerosols that directly alter the atmospheric energy budget. Rapid adjustments to forcings, cooling effects from scattering aerosol, and observational uncertainty can explain why observed global precipitation responses are currently difficult to detect but are expected to emerge and accelerate as warming increases and aerosol forcing diminishes. Precipitation increases with warming are expected to be smaller over land than ocean due to limitations on moisture convergence, exacerbated by feedbacks and affected by rapid adjustments. Thermodynamic increases in atmospheric moisture fluxes amplify wet and dry events, driving an intensification of precipitation extremes. The rate of intensification can deviate from a simple thermodynamic response due to in-storm and larger-scale feedback processes, while changes in large-scale dynamics and catchment characteristics further modulate the frequency of flooding in response to precipitation increases. Changes in atmospheric circulation in response to radiative forcing and evolving surface temperature patterns are capable of dominating water cycle changes in some regions. Moreover, the direct impact of human activities on the water cycle through water abstraction, irrigation, and land use change is already a significant component of regional water cycle change and is expected to further increase in importance as water demand grows with global population.

Keywords: climate change; water cycle; precipitation; land surface; radiative forcing

Introduction

The global water cycle describes a continual circulation of water through Earth's atmosphere, surface and subsurface that taps into the vast stores residing in the ocean, large bodies of ice, and deep within the ground. This cycle also determines smaller,

more transient, yet life-sustaining, stores in rivers and lakes, the upper layers of soil and rock, as well as within animals and vegetation (Fig. 1A). Precipitation over land is strongly dependent on the transport of water vapor from the ocean¹ and the return flow is primarily through rivers (Fig. 1B).

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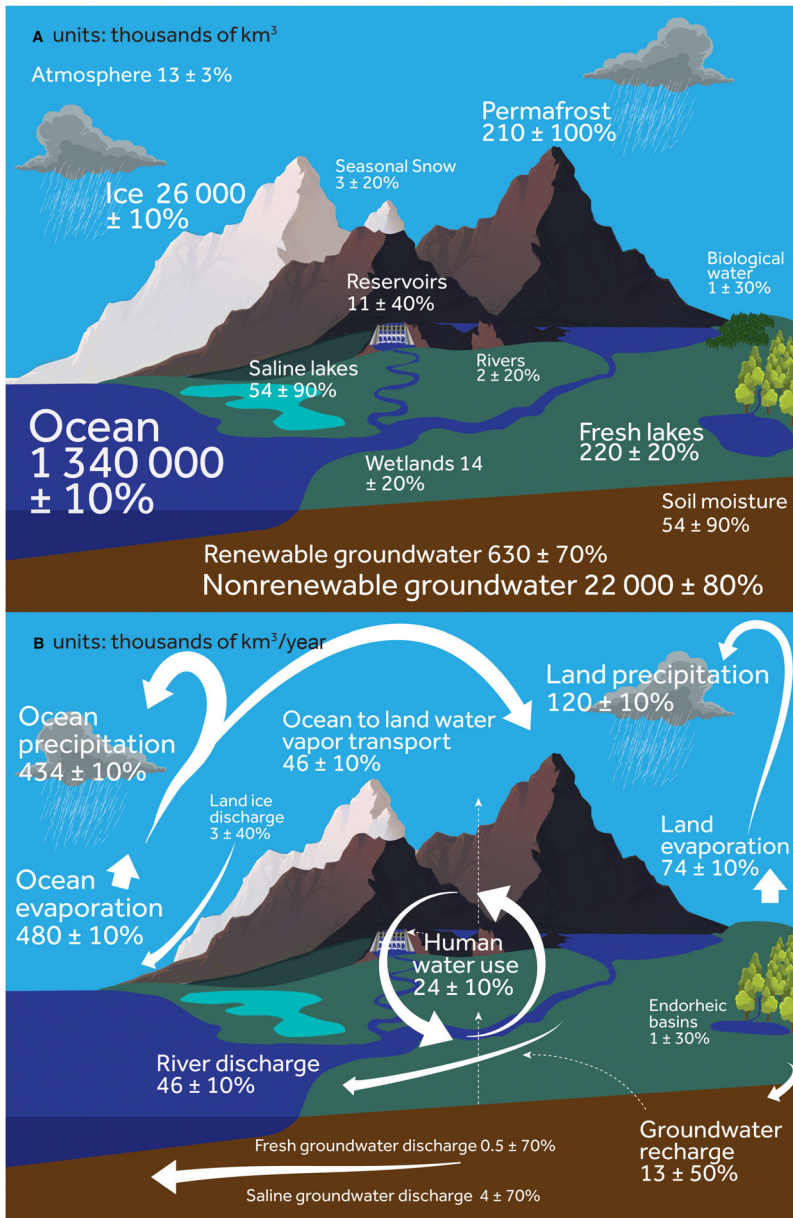


Figure 1. Depiction of the global water cycle: (A) stores (in thousands of km³) and (B) fluxes (thousands of km³ per year) based on previous assessments^{7,14,15} with minor adjustments for fresh groundwater flows¹⁶ and increases in precipitation and evaporation within quoted uncertainty based on observational evidence.¹⁷

The water cycle is influenced by natural variations in the sun and volcanic eruptions, as well as fluctuations internal to the climate system, and there is abundant evidence from the paleoclimate record of substantial past changes.^{2–4} Water cycle changes are increasingly becoming dominated by human activities, indirectly through climatic response to

emissions of greenhouse gases and aerosol particles but also directly from interference with the land surface and the extraction of water from the ground and river systems (Fig. 1B) for agricultural, industrial, and domestic use.^{5–7}

While global mean precipitation changes are determined by Earth's energy balance, regional

changes are dominated by the transport of water vapor and dynamical processes,¹ particularly at scales smaller than ~ 4000 km.⁸ Changes in weather patterns are further determined by altering heating and cooling patterns throughout the atmosphere and across the planet's surface. As the climate changes, these competing constraints operating at global and local scales alter key water cycle characteristics, such as precipitation frequency, intensity, and duration.^{9,10} Future water availability, for use by societies and the ecosystems upon which they depend, is further influenced by increased evaporative demand by the atmosphere,¹¹ but also an increased efficiency of water use by plants in response to elevated CO₂ levels.^{12,13} Societies experience impacts through localized changes in water availability that are controlled by large-scale atmospheric circulation, as well as smaller scale physical processes. At regional to local scales, water cycle changes, therefore, result from the interplay between multiple drivers (CO₂, aerosols, land use change, and human water use). A primary focus here is on reviewing recent advances in understanding how these complex interactions are expected to determine responses in the global water cycle.

Hydrological sensitivity at the global scale

The Clausius–Clapeyron equation is a dominating thermodynamic constraint on atmospheric water vapor. Prevalent increases in atmospheric water vapor with warming¹⁸ drive powerful amplifying climate feedbacks, intensify atmospheric moisture transport and associated heavy precipitation events, and increase atmospheric absorption of sunlight and emission of infrared radiation to the surface that modulate global-scale evaporation and precipitation responses.^{19,20} Simulations and observations confirm a thermodynamic increase in water vapor close to 7%/°C at low altitudes when averaged over global scales.²¹ This sensitivity varies depending on the radiative forcing agent and associated warming pattern: for column integrated water vapor, it ranges from $6.4 \pm 1.5\%/^{\circ}\text{C}^a$ for sulfate aerosol forcing to $9.8 \pm 3.3\%/^{\circ}\text{C}$ for black carbon, based on idealized modeling.²² Changes over global land are below the

thermodynamic response since relative humidity is expected to decrease due to greater land–sea warming contrast²³ that is amplified by land surface feedbacks.²⁴ Multimodel coupled CMIP5 simulations underestimate declining relative humidity observed over global land.^{25,26} This discrepancy also applies to atmosphere-only experiments applying observed sea surface temperature (SST): a single model simulated a -0.05 to $-0.25\%/decade$ trend (1996–2015) compared with an observed estimate of -0.4 to $-0.8\%/decade$.²⁵ It is not clear if this discrepancy is explained by potential deficiencies in representing ocean to land moisture transport,²⁷ land–atmosphere coupling,²⁴ or inhomogeneity of the observational records.²⁸

In contrast to water vapor, global mean evaporation and precipitation are tightly linked to the atmospheric and surface energy budgets rather than the Clausius–Clapeyron equation.^{29,30} Latent heat released through precipitation is balanced by the net atmospheric longwave radiative cooling minus the heating from absorbed sunlight and sensible heat flux from the surface (Fig. 2A). Complementary energetic arguments apply for surface evaporation.^{31,32} The total global mean precipitation response to warming, or apparent hydrological sensitivity (η_a , Fig. 2F) includes fast adjustments that scale with radiative forcing and slow temperature-driven responses to the radiative forcings.^{33–35} The fast response is caused by near-instantaneous changes in the atmospheric energy budget and atmospheric properties (e.g., temperature, clouds, and water vapor; Fig. 2C) in direct response to the radiative effects of a forcing agent.³⁶ A further relatively fast response involves the land–surface temperature (Fig. 2D), which responds more rapidly to radiative forcing than ocean SST.^{35,37} The land surface response depends on the partitioning of increased net surface radiation between latent and sensible heat (and thereby on the land hydrology) as well as the direct response of plants to elevated CO₂.^{24,38} The slower global temperature-dependent precipitation response, or hydrological sensitivity (η , Fig. 2F), is driven by the increased atmospheric radiative cooling rate of a warming atmosphere (Fig. 2E).

The fast and slow responses in global precipitation can be illustrated with idealized experiments as part of the 6th phase of the Coupled Model Inter-comparison Project (CMIP6) in which atmospheric

^a5–95% confidence range is used unless otherwise stated, estimated as $1.645 \times$ standard deviation across models.

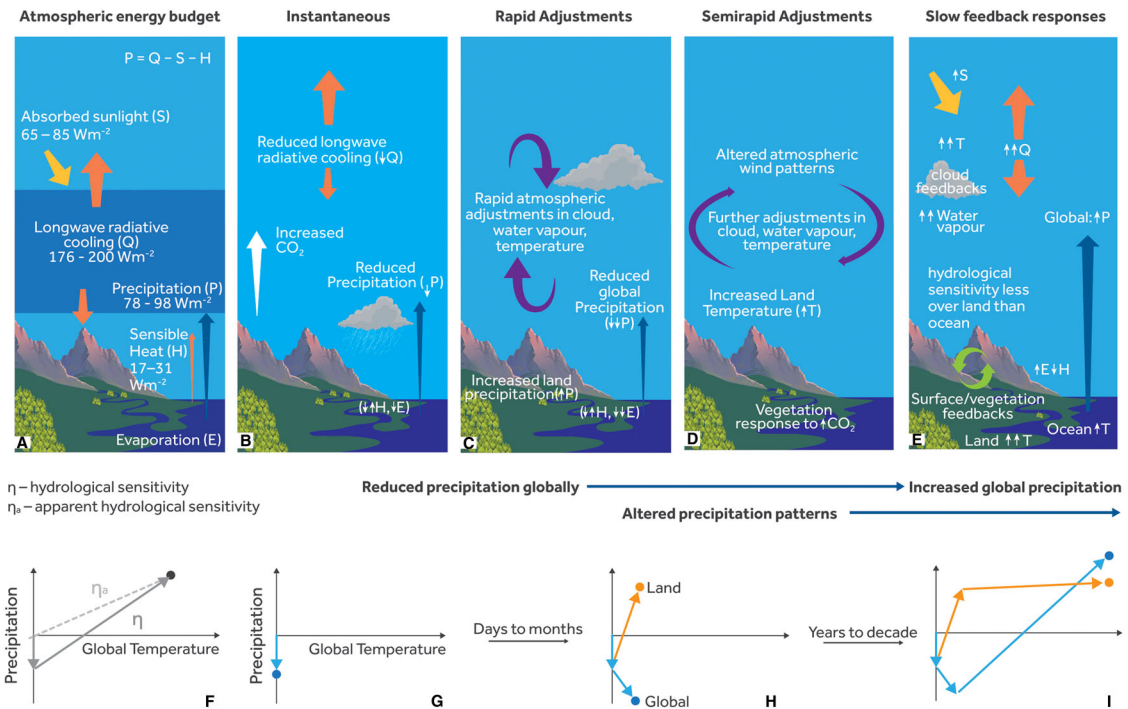


Figure 2. Schematic representation of responses of the atmospheric energy balance and global precipitation to increases in CO₂. The energy budget of the atmosphere (A) responds instantaneously to radiative forcings (B), which leads to rapid atmospheric adjustments (C) and slower semi-rapid adjustments involving the land surface and vegetation that further modify atmospheric circulation patterns (D). As the oceans respond to radiative forcing, longer timescale feedbacks, involving the atmosphere, land, and oceans, alter the surface and atmospheric energy balance, driving increased global evaporation and precipitation (E). This slow response of precipitation to global mean surface temperature is quantified as the hydrological sensitivity, η , while the total precipitation response, including initial fast adjustments, is termed the apparent hydrological sensitivity, η_a (F). The precipitation response over land and ocean develops over time (G–J) with land hydrological sensitivity tending to be suppressed relative to the global mean.

concentrations of CO₂ are instantaneously quadrupled (Fig. 3; simulations listed in Table 1). Global mean precipitation, relative to a preindustrial control, increases linearly with global mean temperature (Fig. 3, black dots and line of best fit) at the rate of 2.7 and 2.3%/K in the two 4 × CO₂ simulations (η , Fig. 2F), consistent with previous estimates of 2.1–3.1%/K.^{39,40} This rate of increase can be understood in terms of radiative transfer that links increased radiative cooling to thermal deepening of the troposphere,³⁵¹ while idealized modeling has recently uncovered the role of surface evaporation as a limiting factor for the atmospheric warming that also determines the magnitude of η .⁴¹ Climate feedbacks also modulate the magnitude of η ,³⁰ and model simulations may underestimate η due to deficiencies in the representation of feedbacks from low-altitude cloud,⁴² which are

linked with hydrological sensitivity through their dependence on temperature lapse rate responses.⁴¹ Uncertainty in the sensitivity of shortwave absorption by atmospheric water vapor to temperature can explain much of the range in simulated hydrological sensitivity,⁴³ although longwave feedbacks also contribute.⁴⁴ Consistency in hydrological sensitivity does, however, disguise contrasting regional responses that are particularly dependent on forcing agent.^{40,44}

The apparent hydrological sensitivity (η_a) is reduced relative to hydrological sensitivity (η) by greenhouse gases and absorbing aerosols, which alter the atmospheric radiation balance, driving rapid adjustments in global precipitation. A rapid adjustment in response to the quadrupling of atmospheric CO₂ concentration is illustrated in Figure 3: following the black regression line back to the

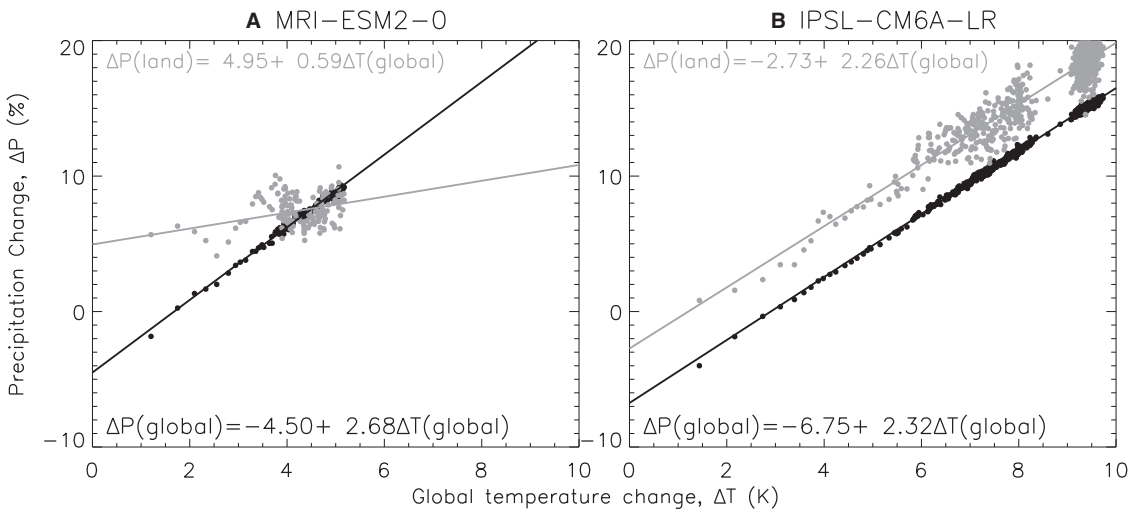


Figure 3. Precipitation changes for global mean (black) and land mean (gray) in response to global mean temperature changes for a $4 \times \text{CO}_2$ experiment relative to a 30-year mean preindustrial control for (A) MRI-ESM2-0 150-year experiment and (B) IPSL-CM6A-LR 900-year experiment (showing the first 300 and last 300 years), where each dot represents 1 year of data.

y-axis implies a decrease in global precipitation before global temperatures have begun increasing in response to the elevated CO_2 levels (-4.5% and -6.8% in the two simulations in Fig. 3). This reflects the rapid adjustments to the atmospheric heating influence of CO_2 radiative forcing, most of which is transferred to the ocean through fast responses in atmospheric vertical motion and circulation. Rapid adjustment effects on precipitation are less certain than the slow responses to surface temperature.^{34,62} The rapid adjustments depend upon how each radiative forcing manifests throughout the atmosphere and surface and explains why the apparent hydrological sensitivity is lower than the hydrological sensitivity for CO_2 forcing (Fig. 2F). Despite uncertainty in the fast precipitation response to radiative forcing, similar spatial patterns are simulated for greenhouse gas, solar, and absorbing aerosol radiative forcings.^{63,64}

Climate drivers that primarily impact the surface rather than atmospheric energy budget initially produce only a small rapid reduction in precipitation. Examples include solar forcing and sulfate aerosol, which produce larger η_a than drivers primarily modulating aspects of the atmospheric energy budget, such as greenhouse gases and absorbing aerosol.^{63,65–67} Thus, global precipitation appears more sensitive to radiative forcing from sulfate aerosols ($2.8 \pm 0.7\%/^\circ\text{C}$, $\eta_a \sim \eta$) than green-

house gases ($1.4 \pm 0.5\%/^\circ\text{C}$, $\eta_a < \eta$), while the response to black carbon aerosol can be negative ($-3.5 \pm 5.0\%/^\circ\text{C}$, $\eta_a < \eta$) due to strong atmospheric solar absorption.⁶³ In four different climate models, the response to a complete removal of present day anthropogenic aerosol emissions was an increase in global mean precipitation ($\eta_a = 1.6\text{--}5.5\%/^\circ\text{C}$), mainly attributed to the removal of sulfate aerosol as opposed to other aerosol species.⁶⁸ η_a also depends on the pattern of aerosol forcing. For example, increased Asian sulfates produce a larger global precipitation response than for comparable aerosol changes over Europe.⁶⁹ The vertical profile of black carbon and ozone influences the magnitude of the fast global precipitation response, yet is more difficult to observe and simulate.^{70–72} The range of apparent hydrological sensitivity obtained from six simulations of the last glacial maximum and preindustrial period ($\eta_a = 1.6\text{--}3.0\%/^\circ\text{C}$) is greater than for a $4 \times \text{CO}_2$ experiment ($\eta_a = 1.3\text{--}2.6\%/^\circ\text{C}$) in which larger CO_2 forcing suppresses precipitation response due to fast adjustments.⁷³ However, thermodynamic constraints on evaporation and contrasting vegetation and land surface states also play a role. A range of fast precipitation adjustments to CO_2 between models is attributed to the response of vegetation, leading to a repartitioning of surface latent and sensible heat fluxes.⁷⁴

Table 1. List of observations and simulations with references

Data set	Period (this study)	Resolution (lat, lon)	References
HadCRUT4v4.6	1979–2018	5° × 5°	45
HadCRUH	1979–2003	5° × 5°	46
SSM/I	1988–2019	0.25° × 0.25°	47
ERA5	1979–2019	0.25° × 0.25°	48
GPCPv2.3	1979–2018	2.5° × 2.5°	49
AMIP6 simulations	1980–2014		
* Preindustrial control	30 years		
* 4 × CO ₂	>150 years		
# Historical	1995–2014		
# SSP2-4.5	2081–2100		
BCC-CSM2-MR		1.125° × 1.125°	50
BCC-ESM1		2.81° × 2.81°	50
CanESM5 [#]		2.8° × 2.8°	51
CESM2		0.94° × 1.25°	52
CNRM-CM6-1		1.4° × 1.4°	53
CNRM-ESM2-1		1.4° × 1.4°	54, 55
GISS-E2-1-G		1.0° × 1.25°	56
IPSL-CM6A-LR*		2.0° × 2.5°	57
MIROC6		1.25° × 2.5°	58
MRI-ESM2-0* [#]		1.406° × 1.406°	59
MRI-ESM2-0* [#]		1.125° × 1.125°	60
UKESM1-0-LL		1.25° × 1.875°	61

All models are used in the AMIP analysis in Figure 4, but only the CanESM5 and MRI-ESM2 historical and SSP2-4.5 experiments are used in Figure 5 (denoted by #) and only the IPSL-CM6A-LR and MRI-ESM2 pre-industrial and 4 × CO₂ experiments are used in Figure 3 (denoted by *).

Hydrological sensitivity is generally suppressed over land (Fig. 2E–I), with a large range ($\eta = 0.8$ – $2.4\%/^{\circ}\text{C}$ for CO₂ doubling experiments) relative to the global mean ($\eta = 2.3$ – $2.7\%/^{\circ}\text{C}$) based on multiple simulations.^{40,44} This is partly explained by the greater warming over land than oceans. Since oceans supply much of the moisture to fuel precipitation over land,^{1,75} the slower ocean warming rate dictates that sufficient moisture cannot be supplied to maintain continental relative humidity,²³ leading to a drying influence that is further amplified by land surface feedbacks.²⁴ A weaker hydrological response over land is important for aridity changes and presents a challenge for attribution of continental precipitation changes to different climate forcings.⁴⁰

The distinct response of water cycle responses over land is illustrated in Figure 3 (gray dots/lines). An implied rapid response in precipitation over land is more positive than the global rapid response in both model simulations. However, one model simulates an initial increase of $\sim 5\%$ over land com-

pared with a 4.5% decrease globally (Fig. 3A), while the other model simulates a decrease of $\sim 3\%$ over land compared with a 7% initial decrease globally (Fig. 3B). The more positive initial precipitation response over land than globally can be explained by rapid land warming, in part from increased surface downwelling longwave radiation. This initially destabilizes the troposphere, strengthening vertical motion, moisture convergence, and precipitation over land in the short term.^{44,76,77} While the hydrological sensitivity over land is similar to the global response in one model (Fig. 3B: $\eta = 2.3\%/^{\circ}\text{C}$), the initial rapid increase in precipitation over land in the other simulation (Fig. 3A) is offset over time through a lower hydrological sensitivity over land ($\eta = 0.6\%/^{\circ}\text{C}$) compared with the global response (Fig. 3A). Continental precipitation increases as a rapid response to CO₂ have been counteracted by past increases in anthropogenic aerosols, which reflect and absorb solar radiation at the expense of surface heating and evaporation of surface moisture.⁷⁸ The precise response depends upon the

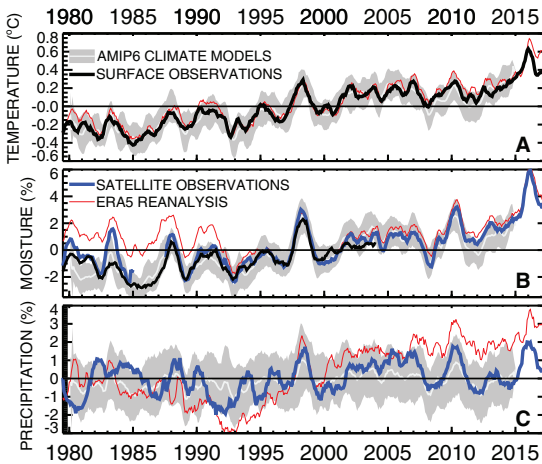


Figure 4. Observed and simulated deseasonalized global mean changes in (A) surface air temperature, (B) column integrated or near surface water vapor, and (C) precipitation with 6-month smoothing and 1994–2000 reference period, including AMIP6 ensemble mean (white line) with shading representing ± 1 standard deviation over 12 models (Table 1) and ERA5 reanalysis.⁴⁸ Observed near surface temperature is from HadCRUTv4.6,⁴⁵ column integrated water vapor is from SSM/I and SMMR satellite data⁴⁷ over ice-free oceans and ERA5 elsewhere, and surface near surface-specific humidity is from HadCRUH⁴⁶ and observed precipitation from GPCP v2.3⁴⁹ and based on previous methods.²¹

aerosol type: sulfate aerosols primarily cool the surface, whereas black carbon aerosols absorb sunlight, heating the atmosphere, and this effect can dominate over the surface cooling effect.⁶³ Recent observations suggest the absorption effects are important in explaining decreases in surface absorbed sunlight that reverse, first in Europe then China, in concert with action to reduce air pollution.⁷⁹ Although aerosol cooling effects have opposed rapid precipitation increases in response to direct CO₂ radiative forcing, these counteracting aerosol effects are expected to diminish with future declining aerosol forcing.^{44,80,81}

Advances in the physical understanding of global precipitation responses can be used to interpret the present day global water cycle changes. Global mean temperature and water vapor are closely coupled (Fig. 4A and B). The linear fit between monthly deseasonalized column integrated water vapor and temperature (1988–2014) is $6.8 \pm 0.4\%/^{\circ}\text{C}$ in the SSM/I satellite-based observations and $7.1 \pm 0.3\%/^{\circ}\text{C}$ in an ensemble of 12 atmosphere-only CMIP6 simulations (AMIP6, which apply observed

SST and sea ice plus realistic radiative forcings; Table 1). This is close to that expected from thermodynamics, assuming small global changes in relative humidity, and is substantially larger than the precipitation sensitivity of $3.2 \pm 0.8\%/^{\circ}\text{C}$ in GPCP observations and $2.0 \pm 0.2\%/^{\circ}\text{C}$ in AMIP6 simulations. These are within the range of η from coupled simulations^{39,40} but are not directly comparable since interannual variability depends on cloud feedbacks specific to ENSO-related changes.⁸² Also shown are the ERA5 reanalysis estimates which, for temperature, show broad consistency with the other data sets. However, the ERA5 depiction of a decrease in water vapor during the early 1990s and larger trends and variability in global precipitation (Fig. 4B and C) are spurious based on the analysis of an earlier reanalysis version,²¹ underlining that global-scale water cycle trends in reanalysis products are not realistic.

Longer term trends are more relevant for expected climate change response, yet are limited by the observing system. Global mean warming of $0.15 \pm 0.01\ ^{\circ}\text{C}/\text{decade}$ and $1.0 \pm 0.1\%/ \text{decade}$ increases in moisture in the observations and AMIP6 simulations (1988–2014) imply a water vapor response of $6.7 \pm 0.3\%/^{\circ}\text{C}$, very close to thermodynamic expectations. Corresponding precipitation trends are not significant at the 95% confidence level in the observations ($0.3 \pm 0.2\%/ \text{decade}$) and AMIP6 simulations ($0.14 \pm 0.06\%/ \text{decade}$), though they are consistent with the role of fast adjustments suppressing hydrological sensitivity in the near term.^{21,83} The implied apparent hydrological sensitivity (η_a) is $2.0 \pm 0.5\%/^{\circ}\text{C}$ in the observations and $0.9 \pm 0.2\%/^{\circ}\text{C}$ in the simulations. Cooling effects of anthropogenic aerosol and rapid adjustments to increases in greenhouse gases and absorbing aerosol reduce global mean precipitation, offsetting increases related to the warming climate. Multidecadal trends in global precipitation for the satellite era are, therefore, expected to be small and difficult to confirm due to observational uncertainty,²¹ and changes in sensible heat flux become significant in determining the precise global hydrological response.⁸³ The warming influence of continued rises in CO₂ concentration, compounded by declining aerosol cooling, is expected to accelerate increases in global precipitation and its extremes as the slow temperature-related responses dominate over rapid atmospheric adjustments to direct radiative forcing effects as

transient climate change progresses.^{21,66,67,83–85} The observational record in Figure 4 is consistent with physical understanding that global mean precipitation increases more slowly than water vapor content per degree of warming. This has important implications since it determines an increase in water vapor lifetime²² and altered precipitation characteristics in terms of regional and seasonal duration, frequency, and intensity.⁸⁶

Thermodynamic constraints on regional precipitation minus evaporation patterns

An important implication of increased atmospheric water vapor with warming (Fig. 4B) is a corresponding intensification of horizontal moisture transport that drives an amplification of existing precipitation minus evaporation (P–E) patterns (Fig. 5). At the regional scale, positive P–E determines fresh water flux from the atmosphere to the surface, while negative P–E signifies a net flux of fresh water into the atmosphere. Atmospheric moisture balance is

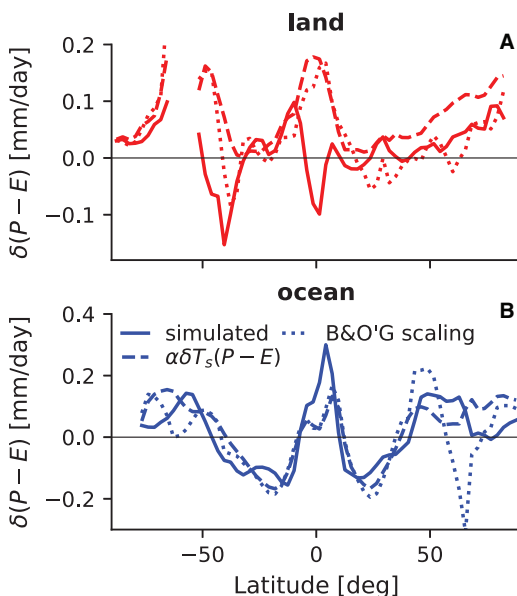


Figure 5. Zonally averaged changes in precipitation minus evaporation $\delta(P-E)$ over (A) land and (B) ocean between the historical (1995–2014) and SSP2-4.5 (2081–2100) simulations (smoothed in latitude using a three-point moving-average filter). The solid lines indicate the simulated changes, which are averages between the CanESM5 and MRI-ESM2-0 models. Dashed lines are a simple thermodynamic scaling, $\alpha\delta T_s(P-E)$, and dotted lines show an extended scaling ($\alpha \approx 0.07 \text{ K}^{-1}$ and δT_s is the change in time-mean local surface air temperature⁸⁷).

achieved primarily by horizontal moisture transport from net evaporative ocean regions into wet convergence zones. At the global scale over the land surface, P–E is positive and approximately balanced by runoff and storage, while over the ocean P–E is negative and approximately balanced by runoff from the land (Fig. 1B), with both factors influencing regional salinity.

A projected amplification of P–E zonal mean patterns over the oceans is explained by the thermodynamic scaling of present day simulated P–E (solid and dashed lines in Fig. 5B). This amplification of zonal mean P–E is corroborated by an observed “fresh get fresher, salty get saltier” salinity response to warming.^{31,88,89} This amplification is moderated by proportionally larger evaporation increases over the subtropical oceans relative to the equatorial convergence zones and weakening of the tropical circulation.⁹⁰ Suppressed evaporation increases over low latitudes ($\sim 1\%/^{\circ}\text{C}$) are partly explained by rapid adjustments to CO_2 increases and uptake of heat by the ocean compared with high latitudes.³² At higher latitudes, evaporation is further increased by the expansion of open water area as sea and lake ice melts with warming.^{91–94} However, ocean stratification due to heating of the upper layers from radiative forcing is identified as a mechanism for amplifying the salinity patterns beyond the responses driven by water cycle changes alone.⁹⁵ Amplified P–E patterns are additionally reduced by atmospheric and ocean circulation changes that alter the locations of the wettest and, therefore, freshest ocean regions. Spatial shifts in atmospheric circulation are thus expected to modify thermodynamic responses locally. This is consistent with paleoclimate evidence showing that mean changes are roughly in agreement with thermodynamic scaling,⁷³ while regional changes are dominated by dynamics.^{96–99} However, ice sheet responses also contribute to regional water cycle change over paleoclimate timescales.^{100–102}

Over land, evaporation is regulated by energy fluxes over wet regions, with atmospheric vapor pressure and aerodynamics playing an important role, but for drier regions evaporation is limited by surface water availability.^{31,103} Changes in P–E over drier continental regions are consequently dominated by precipitation changes³¹ that are strongly determined by alteration in atmospheric circulation. Projected changes in P–E patterns cannot be

simply interpreted as a “wet gets wetter, dry gets drier” response.^{31,87,90,103,104} In a simplistic sense, ocean regions experiencing decreasing P–E cannot meaningfully be described as “dry”³¹, and “dryness” or aridity over land is influenced by potential evaporation as well as precipitation.^{31,104,105} However, a more fundamental objection to “dry gets drier” over land is that P–E is generally positive and balanced by river discharge over multiannual timescales (Fig. 1B), so increased moisture fluxes imply increased P–E with warming.^{31,87,103} It is, however, recognized that P–E may be negative during the tropical dry season or extended dry spells,¹⁰⁶ as ground water is lost to a “more thirsty” atmosphere due to greater evaporative demand^{104,105,107} and exported remotely. Thus, contrasting water cycle responses are expected for wet and dry periods at the seasonal or subseasonal timescale.

Decreases in soil moisture over many subtropical land regions are an expected response to a warming climate.²⁰ Decreases in P–E over land are explained by reductions in relative humidity driven by increased land–ocean warming contrast and spatial gradients in temperature and humidity.^{87,108,109} A simple scaling accounting for these effects captures more closely the simulated responses over subtropical and Northern Hemisphere land (Fig. 5A). Drying over land is further amplified by vegetation responses^{24,108} and drives a reduction in moisture recycling.⁷⁵ The control of soil moisture on evapotranspiration determines feedbacks onto surface climate, which vary across simulations¹¹⁰ and can cause delayed responses over multiple seasons.¹¹¹

The response of vegetation to climate change and increased atmospheric CO₂ concentrations also determines regional P–E, as well as aridity. Depending on their responses, plants may either amplify¹¹² or ameliorate¹¹³ warming impacts on drought at the surface. Plant water use efficiency is determined by the ratio of photosynthesis to transpiration, which in turn is determined by stomatal conductance and vapor pressure deficit. Increased water use efficiency by plants is driven by enhanced photosynthesis and stomatal closure in response to higher CO₂ levels. This can reduce evaporation from vegetated surfaces and exacerbate declining continental relative humidity and precipitation while limiting runoff increases and drying of soils at the root zone.^{12,13,110,113–119} However, increased plant

growth in direct response to elevated CO₂ concentrations that also drives greater tolerance to aridity can counteract increased water use efficiency, thereby offsetting the atmospheric drying, runoff increases, and soil drying effects^{12,13,38,114–116,120,121} Plant physiological responses thereby represent an uncertain component of semi-rapid adjustments to CO₂ forcing (Fig. 2D).

Human activities also directly alter P–E over land. Intensive irrigation increases evapotranspiration and atmospheric water vapor locally. Although increased irrigation efficiency may ensure more water is available to crops, the corresponding reduction in runoff and subsurface recharge may exacerbate hydrologic drought deficits.¹²² Land use change, including deforestation and urbanization, can further alter regional P and E through changes in the surface energy and water balance. Direct human interference with the land surface combined with complex surface feedbacks thereby complicates the expected regional water cycle responses over land. Therefore, while increased moisture transport into wet parts of the atmospheric circulation will amplify P–E patterns globally, the interactions of geography, atmospheric circulation, human activities, and feedbacks involving vegetation and soil moisture lead to a complex regional response over land. However, multiple lines of evidence indicate that the contrast between wet and dry meteorological regimes, seasons, and events will amplify as moisture fluxes increase in a warming climate.^{123–133}

Large-scale responses in atmospheric circulation patterns

Changes in the large-scale atmospheric circulation dominate regional water cycle changes, yet are not as well understood as changes in thermodynamics. Expected large-scale responses in a warming climate are a weakening and broadening of tropical circulation with poleward migration of tropical dry zones and mid-latitude jets.²⁰ Land use change and large-scale irrigation also drive local and remote responses in atmospheric circulation and precipitation by altering the surface energy and moisture balance.^{134–138} Atmospheric circulation responds rapidly to radiative forcing^{40,63,77,139–142} and dominates the spatial pattern of precipitation change in response to different drivers.^{62,77,139,143} Radiative forcings with heterogeneous spatial patterns,

such as ozone and aerosol (particularly related to cloud interactions), drive atmospheric circulation changes through spatially and vertically uneven heating and cooling.^{144–146} These responses are uncertain for aerosol forcing, particularly for black carbon.¹⁴⁷ Robust changes in atmospheric circulation are also driven by slower, evolving patterns of warming, including land–ocean contrasts,^{62,143,148} that are sensitive to model biases.¹⁴⁹

A reduced atmospheric overturning circulation is required to reconcile low-level water vapor increases of $\sim 7\%/^{\circ}\text{C}$ with smaller global precipitation responses of $2\text{--}3\%/^{\circ}\text{C}$, a consequence of thermodynamic and energy budget constraints.²⁰ The slowdown can occur in both the Hadley and Walker circulations, but in most climate models occurs preferentially in the Walker circulation. Paleoclimate simulations and observations support a Walker circulation weakening with warming.¹⁵⁰ However, internal climate variability can temporarily strengthen the Walker circulation over decadal timescales.^{151,152} Although a weaker Walker circulation is associated with El Niño, the associated regional water cycle impacts are not necessarily relevant for climate change responses since the mechanisms driving weakening differ.⁹

There is also a direct link between CO_2 increases and atmospheric circulation response:^{153–155} a rapid $3\text{--}4\%$ slowdown of the large-scale tropical circulation in response to instantaneous quadrupling of CO_2 ¹⁵³ is dominated by reduced tropospheric radiative cooling in subtropical ocean subsidence regions.^{62,77,156} Subsequent surface warming contributes to a slowdown in circulation, the magnitude of which is estimated to reach 12% for a uniform 4°C SST increase, driven by the enhancement of atmospheric static stability through thermodynamic decreases in temperature lapse rate¹⁵³ and an increase in tropopause height.^{20,157} The Hadley cell response is mainly manifest as a widening or poleward shift, partly driven by changes in subtropical baroclinicity and an increase in subtropical static stability.¹⁵⁸

A fundamental component of the Hadley circulation is the Intertropical Convergence Zone (ITCZ), the position, width, and strength of which determine the location and seasonality of the tropical rain belt. Cross-equatorial energy transport is important in determining the mean ITCZ position and both of these attributes display systematic

biases in climate model simulations^{159–164} that can also influence tropical precipitation response to warming.^{165–167} Reduced surface sunlight due to aerosol scattering and absorption that preferentially affects the Northern Hemisphere partially explains a southward shift of the NH tropical edge from the 1950s to the 1980s^{168,169} and the severe drought in the Sahel that peaked in the mid-1980s.^{170,171} Although changes in hemispheric energy imbalance drive relatively small ($<1^{\circ}$ latitude, multidecadal) shifts in the zonally averaged ITCZ position based on observationally constrained simulations,^{172,173} short-term (1–2 years) responses to volcanic eruptions and internal variability can produce more rapid changes.¹⁷⁴ Large shifts in the ITCZ ($>1^{\circ}$ latitude, decades timescale) and regional monsoons are possible following a potential substantial slowdown or collapse of the Atlantic meridional overturning ocean circulation.^{175,176}

Although a dynamical understanding of changes in ITCZ width and strength currently lags understanding of the controls on ITCZ position, energetic and dynamic theories have been developed.^{177–180} Weakening circulation with warming (diagnosed as upward mass transport within the global ITCZ divided by its area) results from a complex interplay between strengthened upward motion in the ITCZ core and weakened updrafts at the edges of the ITCZ.^{161,181} This leads to a drying tendency on the equatorward edges of the ITCZ¹⁷⁷ and a moistening tendency in the ITCZ core: stronger ascent in the ITCZ core amplifies the “wet get wetter” response, while reduced moisture inflow near the ITCZ edges reduces the “wet gets wetter” response relative to the thermodynamic increase in moisture transport. Overall, ITCZ responses have been linked with hemispheric asymmetry in radiative forcing from greenhouse gases and aerosols,^{168,182,183} feedbacks involving clouds,^{184–186} and vertical energy stratification.^{179,187} Changes in the regional tropical rain belt are larger than for the global ITCZ and involve more complex dynamical mechanisms,^{188,189} including monsoons.

Monsoon systems represent an integral component of the seasonal shifts in the tropical rain belt that affect billions of people through the supply of fresh water for agriculture. Onset, retreat, and subseasonal characteristics of monsoons are determined by a complex balance between net energy input by radiative and latent heat fluxes

and the export of moist static energy. This energy export is determined by contrasting surface heat capacity between ocean and land and modified through changes in atmospheric dynamics, tropical tropospheric stability, and land surface properties.^{99,160,190} Thermodynamic intensification of moisture transport increases the intensity and area of monsoon rainfall, but this is offset by a weakening tropical circulation.^{191,192}

Monsoon systems are sensitive to spatially varying radiative forcing relating to anthropogenic aerosol^{168,171,193,194} but also greenhouse gases¹⁸³ and changes in SST patterns^{195,196} that play a strong role by altering cross-equatorial energy transports and land–ocean temperature contrasts. Aerosols affect the monsoon by altering hemispheric temperature gradients and cross-equatorial energy transports but also drive more local changes through altering land–ocean contrasts and changing moisture flux that depend on whether absorbing or scattering aerosol dominate.¹⁹⁷ Reduced surface sunlight due to aerosol increases over land and the oceanic response to reduced cross-equatorial flow can amplify the northward gradient of SST, thereby weakening the Indian monsoon.¹⁴⁸ Although there has been disagreement between paleoclimate and modern observations, physical theory, and numerical simulations of monsoonal changes, many of these discrepancies have been explained by considering regional aspects, such as zonal asymmetries in the circulation, land/ocean differences in surface fluxes, and the character of convective systems.^{98,99,190,198,199}

Poleward expansion of the tropical belt is expected to drive a corresponding shift in mid-latitude storm tracks, yet driving mechanisms differ between hemispheres. Greenhouse gas forcing drives a stronger poleward expansion in the Southern Hemisphere than the Northern Hemisphere. In addition, tropospheric ozone and anthropogenic aerosol forcing contribute to the Northern Hemisphere changes, while an amplification of the Southern Hemisphere response by stratospheric ozone depletion will not apply as ozone levels recover.^{200–203} A thermal gradient between the polar and lower latitude regions that decreases at low altitudes and increases in the upper troposphere as the planet warms is consistent with a strengthening of the winter jet stream in both hemispheres. However, the precise mechanisms are

complex²⁰⁴ and the influence of amplified Arctic warming on mid-latitude regional water cycles is not well understood based on simple physical grounds due to the large number of competing physical processes.^{205–210} Weakening of the Northern Hemisphere summer jet stream is thought to potentially amplify wet and dry extremes through increased persistence of weather types²¹¹ and was linked to reduced precipitation in mid-latitudes based on an early Holocene paleoclimate record.²¹² However, recent analysis of observations and coupled climate simulations show little influence of Arctic warming amplification on mid-latitude climate.^{213,214} Regardless of this uncertainty, thermodynamic increases in moisture and convergence within extratropical cyclones is a robust driver of precipitation increases within mid-high latitude wet events, with implications for more severe flooding.

Changes in characteristics of precipitation and hydrology

Heavy precipitation is expected to become more intense as the planet continues to warm.^{215–217} Increases in low-altitude moisture of around 7%/°C provide a robust baseline expectation for a similar rate of intensification in extreme precipitation, but this is modified by less certain microphysical and dynamical responses^{215,218,219} that are space and timescale dependent.⁸⁶ The response of streamflow and flooding to changing rainfall characteristics is complex (Fig. 6) and there is not a strong relationship between flood hazard and precipitation at the monthly scale.^{220,221} The likelihood of flooding is influenced by snowmelt and antecedent soil moisture^{222–224} that also depend on time and space scales, as well as the nature of the land surface. These complex drivers explain regionally dependent increases and decreases in flooding observed over Europe.^{225,226} Expected drivers of streamflow and flooding are also dependent on direct human intervention, such as river catchment management that can include mismanagement leading to infrastructure failure (e.g., reservoirs), as well as detrimental changes in catchment drainage properties or land stability (e.g., mudslides).

Over mid-latitude regions, the amount and intensity of rainfall within extratropical storms is expected to increase with atmospheric moisture. This is particularly evident for atmospheric rivers: long, narrow bands of intense horizontal moisture

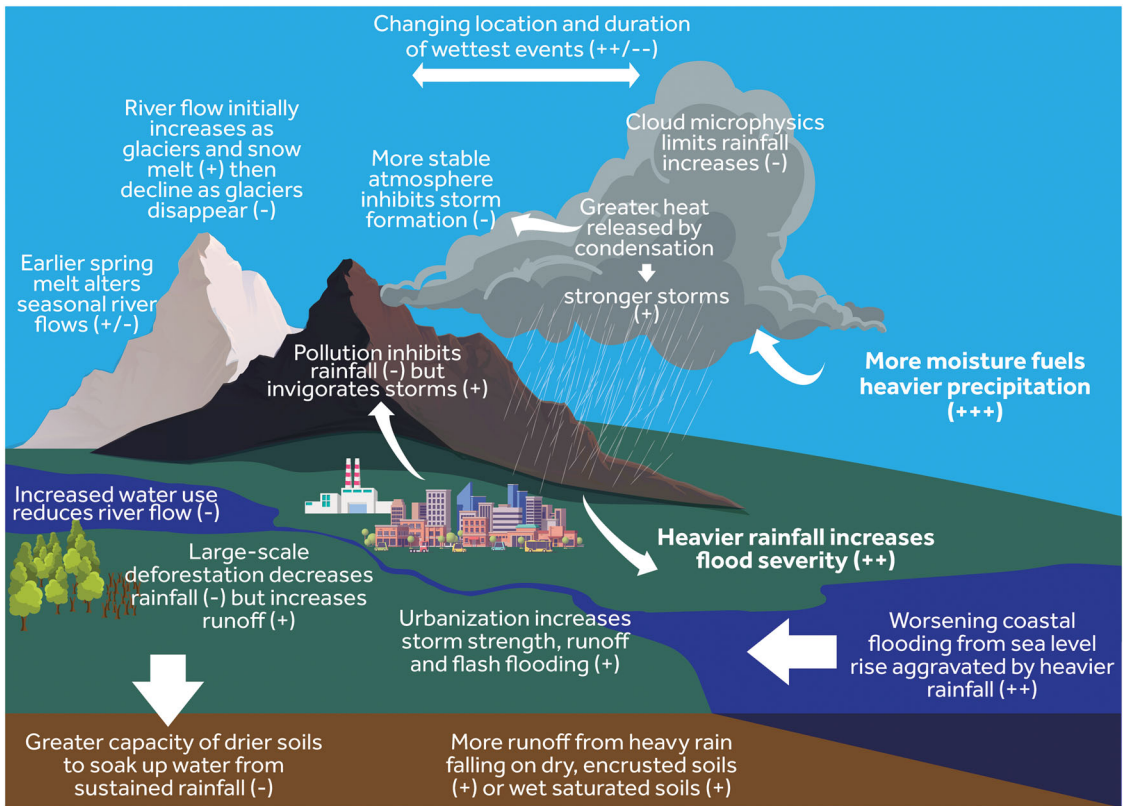


Figure 6. Schematic illustrating factors important in determining changes in heavy precipitation and flooding.

transport within the warm sector of extratropical cyclones^{227,228} that are linked with flooding,^{229–232} changes in terrestrial water storage,²³³ and the mass balance of glaciers and snowpack.^{234–237} Assuming minor changes in dynamical characteristics, it is expected that increased atmospheric moisture flux will intensify atmospheric river events.^{238–241} However, changes in location, orientation, and dynamical aspects relating to wind speed will dominate responses in some regions.

Warming is expected to decrease snowfall globally but could drive increases in intensity regionally, particularly in high latitude winter, since heavy snow tends to occur close to the freezing point,^{242,243} which will migrate poleward, in altitude and seasonally. A shorter snow season can be offset by increased snowfall relating to thermodynamic increases in atmospheric moisture.²⁴⁴ Warming is expected to reduce rain-on-snow melt events at lower altitudes due to declining snow cover but increase these events at higher altitudes

as snow is replaced by rain.^{245,246} Early but less rapid snowmelt is expected from the reduced available radiative energy earlier in the season.²⁴⁷ Earlier and more extensive winter and spring snowmelt²⁴⁸ has been further linked with declining summer and autumn runoff in snow-dominated river basins of mid to high latitudes of the Northern Hemisphere.^{225,249} Increased glacier melt and precipitation are expected to contribute to increasing lake levels, as identified for the inner Tibetan Plateau.²⁵⁰ In a warming climate, glacier runoff is initially expected to increase due to additional melt before decreasing in the longer term as glacier volume shrinks, with peak runoff already achieved for some smaller glaciers.²⁵¹ Changes in the cryosphere thereby drive regional- and seasonal-dependent changes in flooding that may alter in magnitude and even sign over longer timescales.

Increased severity of flooding on larger, more slowly responding rivers is expected as precipitation increases during persistent wet events over

a season. This can occur in mid-latitudes, where blocking patterns continually steer extratropical cyclones across large river catchments, with groundwater flooding also playing a role.^{211,252} Catastrophic floods recorded across Europe and Asia have been linked to persistent atmospheric circulation patterns.^{253–257} Increased atmospheric moisture will amplify the severity of these events when they occur,²⁵⁸ yet changes in occurrence of blocking patterns, stationary waves, and jet stream position depend on multiple drivers and so are not well understood.²⁰⁷ Arctic amplification reduces the low-level latitudinal temperature gradient, which implies a slower or less zonal jet stream and potentially longer duration wet or dry events. However, a stronger temperature gradient in the mid-latitude upper troposphere is expected as the tropical upper troposphere warms and the high-latitude lower stratosphere cools. This potentially drives a stronger jet stream and shorter duration but more intense precipitation associated with the passage of extratropical cyclones, as was found to apply for 30–70°N in CMIP5 projections.²⁵⁹

A weakening tropical circulation is expected to reduce tropical cyclone system speed, thus amplifying thermodynamic intensification of rainfall, though observational evidence supporting this has been questioned.^{260–262} Flooding associated with storm systems can be exacerbated by an increased severity of coastal inundation due to sea level rise.^{263,264} Sensitivity experiments indicate that the most intense rainfall within tropical and extratropical cyclones can increase with warming above the Clausius–Clapeyron rate.^{265,266} There is also observational evidence,^{267–269} supported by simulations,^{270–272} that ingestion of aerosols into tropical cyclones can invigorate the peripheral rain bands and increase the overall area and precipitation of the storm. This occurs at the expense of air converging into the eyewall, thus may decrease the storm's maximum wind speed by up to one class in the Saffire–Simpson scale. However, large-scale cooling from anthropogenic aerosol has been linked with a decreased frequency of tropical storms over the north Atlantic, which reversed at the end of the century as aerosol emissions declined.²⁷³

Increased seasonality in lower latitudes, with more intense wet seasons,^{106,124,129,130,274} will alter seasonal hydrology. Decreases in precursor soil

moisture after more intense dry seasons may increase the timescale over which seasonal rainfall saturates soils and aquifers. Drying of soils can, therefore, reduce the probability of seasonal flooding, while saturated soils associated with more intense wet seasons can increase waterlogging (Fig. 6). Changes in seasonal flood timing in response to climate variability are found to be more sensitive than for rainfall-based metrics. The median change in flood timing over East Africa between El Niño and La Niña of 53 days²⁷⁵ is substantially larger than implied from a rainfall-based estimates of 14 days.²⁷⁶

Increased land–ocean temperature gradients have been linked with more intense precipitation over the Sahel based on satellite data since the 1980s.²⁷⁷ Surface feedbacks, involving soil moisture and vegetation, are also expected to modify regional responses over land,²⁴ including for active to break phase transition over India.^{278,279} The spatial variability in soil moisture has been linked with the timing and location of convective rainfall through altering the partitioning between latent and sensible heating. This has been demonstrated for the Sahel and Europe using satellite data and is not well represented by simulations.^{280–282} Changes in soil moisture and vegetation can, therefore, produce varying effects on rainfall location and intensity.^{283,284} Antecedent soil moisture conditions are an important modulator of flooding but less so for more severe flood events.²²⁴ Defoliation has also been identified as a short-term driver of the regional hydrological cycle with enhanced runoff following a destructive tropical cyclone.²⁸⁵ Increased plant water use efficiency in response to elevated CO₂ concentrations is linked with decreased mean precipitation but increased heavy precipitation days over tropical regions (parts of the Andes, western Amazon, central Africa, and the Maritime Continent) based on modeling experiments.²⁸⁶ More efficient water use by plants can further cause increasing runoff responses to rainfall, particularly for extremes.^{13,287,288}

Precipitation and streamflow are also affected directly by human activities, and water use can offset and even dominate responses to climate change regionally.²⁸⁹ Deforestation can drive increased streamflow as demonstrated by simulations and observations over the Amazon and East Africa,^{290–292} although this can be counterbalanced

by decreases resulting from irrigation.²⁹³ Large-scale forest clearance can also drive reductions in precipitation, which was found to apply for total Amazon deforestation²⁹⁴ but with a substantial range (−38 to +5%) across 44 studies,²⁹⁵ with smaller reductions (−2.3 to −1.3%) estimated from observed Amazon deforestation up to 2010. Small-scale deforestation can actually increase precipitation locally²⁹⁶ and alter storm locations. Altered thermodynamic and aerodynamic properties of the land surface from urbanization can affect precipitation through altered stability and turbulence^{297–299} and are further perturbed through the effect of aerosol pollution on cloud microphysics.³⁰⁰ Urbanization also tends to decrease permeability of the surface, leading to increased surface runoff,³⁰¹ and enhanced urban heat island effects are also known to invigorate convection.^{299,302}

Urban air pollution can invigorate warm base convective storms. The addition of aerosol particles that serve as cloud condensation nuclei (CCN) leads to clouds with more numerous smaller droplets, which are slower to coalesce into raindrops. Therefore, clouds in more polluted air masses need to grow deeper to initiate rain.^{303–305} In clouds with a warm base and depths extending to heights with subzero temperatures, rain suppression increases cloud water that can freeze into large ice hydrometeors and produce heavy rain rates. The added latent heat of freezing can further invigorate the clouds,^{306,308} but simulations indicate this heating may be compensated by changes in latent heat at different cloud altitudes.³⁰⁷ An additional invigoration mechanism, which works mainly in convective tropical clouds with strong coalescence and warm rain, is caused by small aerosol particles (<0.05 μm) that enhance the condensation efficiency of the vapor.³¹³ These cloud invigoration mechanisms redistribute light rainfall from shallow clouds to heavy rainfall from deep clouds. The aerosol convective invigoration effect is nonmonotonic, where the invigoration reverses to weakening at aerosol optical depth greater than ~0.3, though the precise value is dependent on the environmental conditions.^{308–310} This is mainly caused by reduced surface solar heating due to aerosol effects that propels the convection but is also explained by suppression at the cloud edges, which begins to dominate at high aerosol loading.³¹⁰ The magnitude of the ice-forming nuclei effects of aerosols

is poorly known, but likely much smaller than their effects as CCN, except for snow enhancement in shallow orographic clouds.³¹¹ Light-absorbing aerosols, like the microphysical effects of CCN, can redistribute rain intensities from light to heavy. Absorbing aerosol radiative effects increase both instability and convective inhibition, which suppresses the small clouds and enhances the large rain cloud systems.³¹² When the instability is released, often triggered by topographical barriers, intense rainfall and flooding can occur.^{313,314} Such trends were found in India³¹⁵ and eastern China during 1970–2010, and shown to be associated with the increasing amounts of black carbon aerosols there.^{316,317}

Recent advances have been made in understanding the expected changes in subdaily rainfall intensity that can be particularly important in determining flash flooding.³¹⁸ The intensity of convective storms is related to Convective Available Potential Energy, which is expected to increase thermodynamically with warming,^{319,320} although the heaviest rainfall is not necessarily associated with the most intense storms in terms of depth, based on satellite data.³²¹ Intensification can exceed thermodynamic expectations since additional latent heating may invigorate individual storms^{322–328} (Fig. 6), and an increasing height of the tropopause with warming allows the establishment of larger systems³²⁹ that can amplify total storm precipitation.³²⁶ This is corroborated by observed scalings up to three times the rate expected from the Clausius–Clapeyron equation for multiple regions,^{329–332} albeit with low statistical certainty.^{333,334} The relevance of present-day relationships to climate change remains questionable,^{335,336} although is improved by considering scaling with dewpoint temperature, which reduces dependence on dynamical factors.^{329,337,338} Increased frequency of rainfall events above a fixed intensity threshold³³⁹ primarily reflects the less severe precipitation events intensifying above the threshold, so intensification of heavy rainfall in weather systems remains the dominant mechanism.

Intensification of subdaily rainfall is inhibited in regions and seasons where available moisture is limited,³²⁶ and simulations indicate that scaling can depend on time of day.³⁴⁰ However, a fixed threshold temperature above which precipitation is limited by moisture availability is not supported

by recent modeling evidence.^{133,217,326} Enhanced latent heating of the atmosphere by more “juicy” storms can also suppress convection at larger scales due to atmospheric stabilization (Fig. 6), as demonstrated with high-resolution, idealized, and large ensemble modeling studies.^{325,328,341–343} Large eddy simulations demonstrate that stability controls precipitation intensity, moisture convergence governs storm area fraction, while relative humidity determines both intensity and area fraction.³⁴¹ Atmospheric stability is also increased by the direct radiative heating effect from higher concentrations of CO₂³⁴⁴ and aerosol through local effects on the atmospheric energy budget and cloud development. Intensification of short-duration intense rainfall is expected to increase the severity and frequency of flash flooding,^{345,346} and more intense but less frequent storms³²⁸ are also expected to favor runoff and flash flooding at the expense of recharge since a drier surface reduces percolation from intense rain.^{347,348}

Recent modeling evidence shows that increases in convective precipitation extremes are limited by microphysical processes involving droplet/ice fall speeds.^{345,349} Although instantaneous precipitation extremes are sensitive to microphysical processes, daily extremes are determined more by the degree of convective aggregation in one comparison of idealized model simulations.³⁵⁰ Thus, regional processes and their impact on dynamical responses are crucial in determining how regional precipitation intensity and hydrology respond to climate change (Fig. 6). Thermodynamic factors are, however, crucial in determining an intensification of heavy rainfall and associated flooding when extreme events occur.

Conclusions

Based on the physical understanding of thermodynamic processes, corroborated by observations and comprehensive simulations, the global water cycle is expected to intensify with warming in terms of moisture fluxes within the atmosphere and exchanges with the land and ocean surface. This intensification will be offset by a weakening tropical circulation in response to changes in the global energy balance and regional temperature gradients. It is well understood that thermodynamic increases in low-altitude water vapor of about 7%/°C are larger than the 2–3%/°C increases in global evap-

oration and precipitation that are driven by Earth’s evolving energy balance in response to warming. The slowing of atmospheric circulation is required to reconcile these contrasting responses that also imply an increased water vapor residence time. Combined with more intense fluxes of moisture, this is expected to manifest as a region- and season-dependent shift in the distribution of precipitation characteristics, such as intensity, frequency, and duration. Increases in aerosols offset some of the warming effects that drive the intensification of the hydrological cycle, but this depends on the mix of aerosol species and there are strong regional variations. Regionally, more intense moisture fluxes will drive an amplification of wet and dry seasons and weather events, with the possibility of increased duration or persistence driven by tropical circulation weakening. However, regional increases and decreases in precipitation or aridity are expected to be dominated by spatial shifts in atmospheric wind patterns in many regions that alter the location of the wettest and driest parts of the global circulation, yet are less certain than thermodynamic drivers. Local-scale effects are further modulated by land surface feedbacks and vegetation responses to rising concentrations of CO₂, as well as direct human interference with the water cycle through water use and land use change.

Recent advances in refining how the water cycle is expected to respond to continued emissions of greenhouse gases and aerosol are as follows:

- Understanding of how global precipitation and evaporation increase as the planet warms has strengthened based on idealized modeling. Precipitation and atmospheric circulation respond rapidly to different radiative forcing agents but with moderate uncertainty. There is greater certainty in the global response to the slower evolving warming patterns.
- It is now recognized that cooling from sulfate aerosol and atmospheric heating due to rising concentrations of absorbing aerosol has suppressed global precipitation increases over recent decades. However, the dominating greenhouse gas warming influence is expected to drive substantial future global precipitation increases closer to the hydrological sensitivity of 2–3%/°C, with an additional, temporary

acceleration of precipitation increases due to declining aerosol forcing.

- Hydrological sensitivity over land is suppressed relative to the global mean and this has been related to land–ocean warming contrast and surface feedbacks. However, simulated responses are uncertain and do not fully capture the observed magnitude of continental relative humidity decline.
- There is further evidence that amplification of precipitation minus evaporation patterns is robust over the ocean. Understanding of responses over land has been refined beyond an inaccurate “wet get wetter, dry get drier” response. Now recognized as important are regional thermodynamic responses and feedbacks and how aridity or dryness depends on which aspects of the atmosphere, soil, or vegetation are the primary focus.
- There is increasing evidence that the water cycle is intensifying with increased moisture fluxes driving heavier rainfall. Amplified fresh water transport and exchanges between the atmosphere and surface are intensifying wet and dry seasons or weather events.
- Although atmospheric circulation responses are less certain than thermodynamic drivers, evidence for a weaker Walker circulation in a warmer climate has expanded. There is, however, recognition that internal variability can lead to temporary strengthening over a decadal timescale.
- Thermodynamic amplification of monsoon intensity is offset by a weakening tropical circulation, but additional suppression of monsoon precipitation due to reduced solar heating from aerosols is expected to reverse as aerosol emissions decline.
- There have been advances in understanding how hemispheric asymmetries in radiative forcing impact the tropical rain belt, with Northern Hemisphere cooling from sulfate aerosol implicated in a southward shift in the ITCZ associated with the 1980s Sahel drought. Greenhouse gas forcing is now thought to have contributed to the recovery in Sahel rainfall through intensification of the Sahara heat low.
- Recent evidence indicates a limited role for Arctic amplification of warming and the rapid reduction in sea ice area in modifying mid-latitude weather patterns, including the frequency of persistent jet stream position that can favor flooding or drought.
- There is a growing appreciation for the role of vegetation and land surface feedbacks on water cycle responses. Understanding of the direct response of plants to elevated CO₂ concentrations has also advanced. Reduced stomatal conductance increases water use efficiency, thereby reducing transpiration, atmospheric humidity, and local precipitation. This can limit drying of soils and increased streamflow induced by climate change. However, increased photosynthesis and plant growth is also capable of counteracting the effects of increased water use efficiency in some regions for species that are not subject to severe water limitation.
- The role of atmospheric rivers in determining regional water stores in the ground and as snow or ice, as well as extreme rainfall and flooding, is increasingly recognized.
- There is a greater appreciation of the seasonal complexity in water cycle changes as wet and dry periods intensify, but the timing and characteristics of wet seasons, melt events, and streamflow evolve over time.
- Nonlinear changes in streamflow over multidecadal timescales are expected in some regions as accelerated glacier melt is followed by declining glacier volume. This can result in a peak in river discharge that has already been passed in some catchments.
- There have been advances in understanding responses of subdaily precipitation, including the possibility of storm invigoration through enhanced latent heating within storms but convective inhibition operating at larger scales as heat release stabilizes the atmosphere. Responses are thereby dependent on time and space scale, though uncertainty remains in modeling storm systems and their aggregation.
- There have been some advances in identifying the role of aerosols in cloud development through initial suppression of precipitation but deepening of clouds that drive convective invigoration in tropical clouds. The observed shift of rain intensities from low to high can in some cases be related to the

combined microphysical and radiative effects of aerosols, suppressing the small and shallow convective clouds and enhancing the large and deep clouds.

- The role of land–sea temperature gradients, surface feedbacks involving soil moisture and vegetation, as well as deforestation are important in determining the location and intensity convective storms, while questions remain as to their representation in models.
- There is not a simple relation between rainfall intensification and flooding, though evidence has strengthened that the most severe flooding situations will worsen, especially for smaller catchments and urban environments, and can compound increased coastal inundation from sea level rise.
- There is now a greater appreciation for the direct impact of human activity on the water cycle through extraction of water from the ground and river systems for irrigation and industrial or domestic use, as well as how land use change can alter the surface energy and water balances; for example, large-scale deforestation is linked with increased streamflow but also altered wind patterns and reduced precipitation and humidity locally.

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Author contributions

R.P.A. conducted the analysis and lead writing of the paper, M.B. produced Figure 5, and all authors contributed to the assessment of the literature and writing of the manuscript.

Competing interests

The authors declare no competing interests.

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