Water cycle changes

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Chapter 8: Water cycle changes

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Executive Summary

This chapter assesses multiple lines of evidence to evaluate past, present and future changes in the global water cycle. It complements material in Chapters 2, 3, and 4 on observed and projected changes in the water cycle, and Chapters 10 and 11 on regional climate change and extreme events. The assessment includes the physical basis for water cycle changes, observed changes in the water cycle and attribution of their causes, future projections and related key uncertainties, and the potential for abrupt change. Paleoclimate evidence, observations, reanalyses and global and regional model simulations are considered. The assessment shows widespread, non-uniform human-caused alterations of the water cycle, which have been obscured by a competition between different drivers across the 20th century that will be increasingly dominated by greenhouse gas (GHG) forcing at the global scale.

Physical Basis for Water Cycle Changes

Modifications of Earth's energy budget by anthropogenic radiative forcings drive substantial and widespread changes in the global water cycle. There is *high confidence* that global mean precipitation and evaporation increase with global warming, but the estimated rate is model-dependent (*very likely* range of 1–3% per 1°C). The global increase in precipitation is determined by a robust response to global mean surface air temperature (*very likely* 2–3% per 1°C) that is partly offset by fast atmospheric adjustments to atmospheric heating by GHGs and aerosols. The overall effect of anthropogenic aerosols is to reduce global precipitation and alter large-scale atmospheric circulation patterns through their well-understood surface radiative cooling effect (*high confidence*). Land-use and land-cover changes also drive regional water cycle changes through their influence on surface water and energy budgets (*high confidence*). {8.2.1, 8.2.3.4, 8.2.2.2, Box 8.1}

A warmer climate increases moisture transport into weather systems, which, on average, makes wet seasons and events wetter (*high confidence*). An increase in near-surface atmospheric water holding capacity of about 7% per 1°C of warming explains a similar magnitude of intensification of heavy precipitation events (from sub-daily up to seasonal time scales) that increases the severity of flood hazards when these extremes occur (*high confidence*). The severity of very wet and very dry events increases in a warming climate (*high confidence*), but changes in atmospheric circulation patterns alter where and how often these extremes occur with substantial regional differences and seasonal contrasts. The slowdown of tropical circulation with global warming partly offsets the warming-induced strengthening of precipitation in monsoon regions (*high confidence*). {8.2.2, 8.2.3, 8.3.1.7, 8.4.1, 8.5.1}

Warming over land drives an increase in atmospheric evaporative demand and the severity of droughts (*high confidence*). Greater warming over land than over the ocean alters atmospheric circulation patterns and, on average, reduces continental near-surface relative humidity, which contributes to regional drying (*high confidence*). Increasing atmospheric CO_2 concentrations increase plant growth and water-use efficiency, but there is *low confidence* in how these factors drive regional water cycle changes. {8.2.2, 8.2.3}

Causes of Observed Changes

Human-caused climate change has driven detectable changes in the global water cycle since the mid-20th century (*high confidence*). Global warming has contributed to an overall increase in atmospheric moisture and precipitation intensity (*high confidence*), increased terrestrial evapotranspiration (*medium confidence*), influenced global patterns in aridity (*very likely*), and enhanced contrasts in surface salinity and precipitation minus evaporation patterns over the oceans (*high confidence*). {3.4.2, 3.4.3, 3.5.2, 8.3.1, 9.2.2}

Greenhouse gas forcing has driven increased contrasts in precipitation amounts between wet and dry seasons and weather regimes over tropical land areas (*medium confidence*), with a detectable precipitation increase in the northern high latitudes (*high confidence*). GHG forcing has also contributed to drying in dry summer climates, including the Mediterranean, southwestern Australia, southwestern South America, South Africa, and western North America (*medium to high confidence*). Earlier onset of spring snowmelt and increased melting of glaciers have already contributed to seasonal changes in streamflow in high-latitude and low-elevation mountain catchments (*high confidence*). {Box 8.2, 8.2.2.1, 8.3.1, 3.3.2, 3.3.3, 3.5.2}

Anthropogenic aerosols have driven detectable large-scale water cycle changes since at least the mid-20th century (*high confidence*). Shifts in the tropical rain belt are associated with the inter-hemispheric temperature response to the time-evolving radiative influence of anthropogenic aerosols and the ongoing warming influence of GHGs (*high confidence*). Cooling in the Northern Hemisphere by sulphate aerosols explained a southward shift in the tropical rain belt and contributed to the Sahel drought from the 1970s to the 1980s (*high confidence*), subsequent recovery from which has been linked with GHG warming (*medium confidence*). Observed changes in regional monsoon precipitation, especially over South Asia, East Asia and West Africa, have been limited over much of the 20th century due to increases driven by warming from GHGs being counteracted by decreases due to cooling from anthropogenic aerosols (*high confidence*). {8.3.1.3, 8.3.2.4, Box 8.1}

Land-use change and water extraction for irrigation have influenced local and regional responses in the water cycle (*high confidence*). Large-scale deforestation has likely decreased evapotranspiration and precipitation and increased runoff over the deforested regions. Urbanization has increased local precipitation (*medium confidence*) and runoff intensity (*high confidence*). Increased precipitation intensities have enhanced groundwater recharge, most notably in tropical regions (*medium confidence*). There is *high confidence* that groundwater depletion has occurred since at least the start of the 21st century as a consequence of groundwater withdrawals for irrigation in agricultural areas in drylands (e.g., the United States southern High Plains, California Central Valley, North China Plain, and northwest India). {8.2.3.4, 8.3.1.7, Box 10.3, FAQ8.1}

Southern Hemisphere storm tracks and associated precipitation have shifted polewards since the 1970s, especially in the austral summer and autumn (*high confidence*). It is *very likely* that these changes are associated with a positive trend in the Southern Annular Mode, related to both stratospheric ozone depletion and GHG increases. There is *medium confidence* that the recent observed expansion of the Hadley Circulation was caused by GHG forcing, especially in the Southern Hemisphere, but there is only *low confidence* in how it influences the drying of subtropical land areas. {8.2.2, 8.3.2, 3.3.3}

Future Water Cycle Changes

Without large-scale reduction in greenhouse gas emissions, global warming is projected to cause substantial changes in the water cycle at both global and regional scales (*high confidence*). Global annual precipitation over land is projected to increase on average by 2.4 [–0.2 to 4.7] % (*very likely* range) in the SSP1-1.9 low-emission scenario and by 8.3 [0.9 to 12.9] % in the SSP5-8.5 high-emission scenario by 2081–2100, relative to 1995–2014. It is *virtually certain* that evaporation will increase over the oceans and *very likely* that evapotranspiration will increase over land with regional exceptions in drying areas. There is *low confidence* in the sign and magnitude of projected changes in global land runoff in all Shared-socioeconomic Pathway scenarios. Projected increases in precipitation amount and intensity will be associated with increased runoff in the northern high latitudes (*high confidence*). There is *high confidence* that mountain glaciers will diminish in all regions and that seasonal snow cover duration will generally decrease. Runoff from small glaciers will typically decrease through loss of ice mass, while runoff from large glaciers is *likely* to increase with increasing global warming until glacier mass becomes depleted (*high confidence*). {4.5.1, 8.4.1}

Increased evapotranspiration due to growing atmospheric water demand will decrease soil moisture over the Mediterranean, southwestern North America, south Africa, southwestern South America, and southwestern Australia (*high confidence*). The total land area subject to increasing drought frequency and severity will expand (*high confidence*), and in the Mediterranean, southwestern South America, and western North America, future aridification will far exceed the magnitude of change seen in the last millennium (*high confidence*). Some tropical regions are also projected to experience increased aridity, including the Amazon basin and Central America (*high confidence*). {8.4.1}

Water cycle variability and extremes are projected to increase faster than average changes in most regions of the world and under all emission scenarios (*high confidence*). In the tropics and in the extratropics of both hemispheres during summer/warm season, interannual variability of precipitation and runoff over land is projected to increase at a faster rate than changes in seasonal mean precipitation amount (*medium confidence*). It is *very likely* that rainfall variability related to the El Niño–Southern Oscillation will be amplified by the end of the 21st century. Sub-seasonal precipitation variability is also projected to increase, with fewer rainy days but increased daily mean precipitation intensity over many land regions (*high confidence*). Precipitation extremes will increase in almost all regions (*high confidence*), even where seasonal mean precipitation is projected to decrease (*medium confidence*). There is *high confidence* that heavy precipitation events associated with both tropical and extratropical cyclones will intensify. {4.5.1.4, 4.5.3.2, 8.2.3.2, 8.4.1, 8.4.2, 8.5.2, 11.7.1.5}

There are contrasting projections in monsoon precipitation, with increases in more regions than decreases (*medium confidence*). Summer monsoon precipitation is projected to increase for the South, Southeast and East Asian monsoon domains, while North American monsoon precipitation is projected to decrease (*medium confidence*). West African monsoon precipitation is projected to increase over the Central Sahel and decrease over the far western Sahel (*medium confidence*). There is *low confidence* in projected precipitation changes in the South American and Australian monsoons (for both magnitude and sign). There is *high confidence* that the monsoon season will be delayed in North and South America and *medium confidence* that it will be delayed in the Sahel. {8.2.2, 8.4.2.4}

Precipitation associated with extratropical storms and atmospheric rivers will increase in the future in most regions (*high confidence*). A continued poleward shift of storm tracks in the Southern Hemisphere (*likely*) and the North Pacific (*medium confidence*) will lead to similar shifts in annual or seasonal precipitation. There is *low confidence* in projections of blocking and stationary waves and therefore their influence on precipitation for almost all regions. {8.4.2}

The seasonality of precipitation, water availability and streamflow will increase with global warming over the Amazon (*medium confidence*) and in the subtropics, especially in the Mediterranean and southern Africa (*high confidence*). The annual contrast between the wettest and driest month of the year is *likely* to increase by 3–5% per 1°C in most monsoon regions in terms of precipitation, precipitation minus evaporation, and runoff (*medium confidence*). There is *high confidence* in an earlier onset in spring snowmelt, with higher peak flows at the expense of summer flows in snow-dominated regions globally, but *medium confidence* that reduced snow volume in lower-latitude regions will reduce runoff from snowmelt. {8.2.2, Box 8.2, 8.4.1.7, 8.4.2.4}

Confidence in Projections, Non-Linear Responses and the Potential for Abrupt Changes

Representation of key physical processes has improved in global climate models (GCMs), but they are still limited in their ability to simulate all aspects of the present-day water cycle and to agree on future changes (*high confidence*). Climate change studies benefit from sampling the full distribution of model outputs when considering future projections at regional scales. Increasing horizontal resolution in GCMs improves the representation of small-scale features and the statistics of daily precipitation (*high confidence*). High-resolution climate and hydrological models provide a better representation of land surfaces, including topography, vegetation and land-use change, which improve the accuracy of simulations of regional changes in the water cycle (*high confidence*). There is *high confidence* in the potential added value of regional climate models but only *medium confidence* that this potential is currently realized. {8.5.1}

Natural climate variability will continue to be a major source of uncertainty in near-term (2021–2040) water cycle projections (*high confidence*). Decadal predictions of water cycle changes should be considered with *low confidence* in most land areas because the internal variability of precipitation is difficult to predict and can offset or amplify the forced water cycle response. Water cycle changes that have already emerged from natural variability will become more pronounced in the near term, but the occurrence of volcanic eruptions (either single large events or clustered smaller ones) can alter the water cycle for several years, decreasing global mean land precipitation and altering monsoon circulation (*high confidence*). {8.5.2, CCB4.1}

Continued global warming will further amplify GHG-induced changes in large-scale atmospheric circulation and precipitation patterns (*high confidence*), but in some cases regional water cycle changes are not linearly related to global warming. Non-linear water cycle responses are explained by the interaction of multiple drivers, feedbacks and time scales (*high confidence*). Nonlinear responses of regional runoff, groundwater recharge and water scarcity highlight the limitations of simple pattern-scaling techniques (*medium confidence*). Water resources fed by melting glaciers are particularly exposed to nonlinear responses (*high confidence*). {8.5.3}

Abrupt human-caused changes to the water cycle cannot be excluded. There is evidence of abrupt change in some high-emission scenarios, but there is no overall consistency regarding the magnitude and timing of such changes. Positive land-surface feedbacks, including vegetation and dust, can contribute to abrupt changes in aridity, but there is only *low confidence* that such changes will occur during the 21st century. Continued Amazon deforestation, combined with a warming climate, raises the probability that this ecosystem will cross a tipping point into a dry state during the 21st century (*low confidence*). The paleoclimate record shows that a collapse in the Atlantic Meridional Overturning Circulation (AMOC) causes abrupt shifts in the water cycle (*high confidence*), such as a southward shift in the tropical rain belt, weakening of the African and Asian monsoons and strengthening of Southern Hemisphere monsoons. There is *medium confidence* that AMOC will not collapse before 2100, but should it collapse, it is *very likely* that there would be abrupt changes in the water cycle. {8.6.1, 8.6.2}

Solar radiation modification (SRM) could drive abrupt changes in the water cycle (*high confidence*). It is *very likely* that abrupt water cycle changes will occur if SRM techniques are implemented rapidly or terminated abruptly. The impact of SRM is spatially heterogeneous (*high confidence*), will not fully mitigate the GHG-forced water cycle changes (*medium confidence*), and can affect different regions in potentially disruptive ways (*low confidence*). {8.6.3}

8.1 Introduction

8.1.1 Scope and overview

8.1.1.1 Importance of water for human societies and ecosystems

Water is vital to all life on Earth. 71% of the Earth is covered by water, with saline ocean water accounting for around 96.6% of total water availability (Figure 8.1). Terrestrial freshwater only represents about 1.8% of all water on Earth and the remainder (1.6%) is primarily made up of saline groundwater and saline lakes (Durack, 2015; Abbott et al., 2019). Ice sheets, glaciers and snow pack account for approximately 97% of freshwater resources, with less than 3% of freshwater considered easily accessible and available for essential ecosystem functioning and human society's water resource needs (Durack, 2015; Abbott et al., 2019). This very small fraction of freshwater represents a total volume of about 835 thousand km^3, mostly contained in groundwater (630 thousand km^3), the remaining 205 thousand km^3 being stored in lakes, rivers, wetlands and soils (Abbott et al., 2019). Although the natural cycling rate of this amount is theoretically enough to meet global human and ecosystem needs, there are large geographical and seasonal differences that influence the availability of freshwater to meet regional demands.

Freshwater is the most essential natural resource on the planet (Mekonnen and Hoekstra, 2016; Djehdian et al., 2019) and underpins almost all Sustainable Development Goals (SDGs), which require access to adequate and safe resources for drinking and sanitation (SDG 6) and many other purposes. Freshwater supports a range of human activities from irrigation to industrial processes including the generation of hydroelectricity and the cooling of thermoelectric power plants (Bates et al., 2008; Schewe et al., 2014). These activities require sufficient quantities of freshwater that can be drawn from rivers, lakes, groundwater stores, and in some cases, desalinated sea water (Schewe et al., 2014). Recent estimates of global water pools and fluxes suggest that half of global river discharge is redistributed each year by human water use (Abbott et al., 2019). This emphasises the need to consider both anthropogenic climate change and direct human influences, such as population increase or migration, economic development, urbanization, and land use change, when planning water-related mitigation or adaptation strategies (Jiménez Cisneros et al., 2014).

Water scarcity occurs when there are insufficient freshwater resources to meet water demands, although water problems may also arise from water quality issues or from economic and institutional barriers (WGII Chapter 4). This affects the preservation of environmental flows that ultimately influence ecosystem functioning and services (Schewe et al., 2014; Mekonnen and Hoekstra, 2016; Djehdian et al., 2019). As such, water availability is a major constraint on human society’s ability to meet the future food and energy needs of a growing population (D’Odorico et al., 2018). Water plays a key role in the production of energy, including hydroelectricity, bioenergy, and the extraction of unconventional fossil fuels (Schewe et al., 2014; D’Odorico et al., 2018; Djehdian et al., 2019). These dependencies have resulted in increasing competition for water between the food and energy sectors. Pressures on this ‘food-energy-water nexus’ are further compounded by increasing globalization, which can transfer large-scale water demands to other regions of the world, raising serious concerns about local food and water security in regions that are highly dependent on agricultural exports or imports (D’Odorico et al., 2018).

The consequences of climate change on terrestrial ecosystems and human societies are primarily experienced through changes to the global water cycle (Jiménez Cisneros et al., 2014). Changes in the quantity and seasonality of water due to climate change have long been recognized by IPCC and global development agencies as heavily influencing the food security and economic prosperity of many countries, particularly in the arid and semi-arid areas of the world including Asia, Africa, Australia, Latin America, the Mediterranean, and small island developing states (Bates et al., 2008; Schewe et al., 2014; Mekonnen and Hoekstra, 2016). Having too much or too little water increases the likelihood of flooding and drought, as precipitation variability increases in a warming climate (Stocker et al., 2013; Hoegh-Guldberg et al., 2019). Climate change poses a threat to both regional water availability and global water security. Changes in precipitation and glacier runoff and snowmelt influence other hydroclimate variables like surface and subsurface runoff, and groundwater recharge, which are critical to the water, food and energy security of many regions (Oki and Kanae, 2006; Jiménez Cisneros et al., 2014; Schewe et al., 2014; Mekonnen and

Hoekstra, 2016).

Currently, around four billion people live under conditions of severe freshwater scarcity for at least one month of the year, with half a billion people in the world facing severe water scarcity all year round (Mekonnen and Hoekstra, 2016). AR5 WGII reported that approximately 80% of the world's population already suffers from high levels of threat to water security (Jiménez Cisneros et al., 2014). Given the vulnerability of the planet's freshwater resources and the role of climate change in intensifying adverse impacts on human societies and ecosystems (IPCC, 2018; Hoegh-Guldberg et al., 2019), this chapter evaluates advances in the theoretical, observational and model based understanding of the global water cycle made since AR5 (IPCC, 2013) and AR6 Special Reports.

8.1.1.2 Overview of the global water cycle in the climate system

As shown in Figure 8.1, the global water cycle is the continuous, naturally occurring movement of water through the climate system from its liquid, solid and gaseous forms among reservoirs of the ocean, atmosphere, cryosphere and land (Stocker et al., 2013). In the atmosphere, water primarily occurs as a gas (water vapour), but it is also present as ice and liquid water within clouds where it substantially affects Earth's energy balance (Section 7.4.2.2 and 7.4.2.4). The water cycle primarily involves the evaporation[1] and precipitation of moisture at the Earth's surface including transpiration associated with biological processes. Water that falls on land as precipitation, supplying soil moisture, groundwater recharge, and river flows, was once evaporated from the ocean or sublimated from ice-covered regions before being transported through the atmosphere as water vapour, or in some areas was generated over land through evapotranspiration (Gimeno et al., 2010; van der Ent and Savenije, 2013). In addition, the net flux of atmospheric and continental freshwater is a key driver of sea surface salinity, which in turn influences the density and circulation of the ocean (Chapter 9).

Understanding the interactions between the water and energy cycles is one of the four core projects of the World Climate Research Programme (WCRP). Latent heat fluxes, released by condensation of atmospheric water vapour and absorbed by evaporative processes, are critical to driving the circulation of the atmosphere on scales ranging from individual thunderstorm cells to the global circulation of the atmosphere (Stocker et al., 2013; Miralles et al., 2019). Water vapour is the most important gaseous absorber in the Earth's atmosphere, playing a key role in the Earth's radiative budget (Schneider et al., 2010). As atmospheric water vapour content increases with temperature, it has a considerable influence on climate change (Section 7.4.2.2). Additionally, a small fraction of the atmospheric water content is liquid or solid and has a major effect on both solar and longwave radiative fluxes, from the Earth's surface to the top of the atmosphere. The cloud response to anthropogenic radiative forcings, both in the tropics and in the extratropics (Zelinka et al., 2020), is therefore also crucial for understanding climate change (Section 7.4.2.4).

The terrestrial water and carbon cycles are also strongly coupled (Cross-Chapter Box5.1). As atmospheric carbon dioxide (CO_2) concentration increases, the physical environment in which plants grow is altered, including the availability of soil moisture necessary for plants' CO_2 uptake and, potentially, the effectiveness of carbon dioxide removal techniques to mitigate climate change (Section 5.6.2.1.2). Rising surface CO_2 concentrations also modify stomatal (small pores at the leaf surface) regulation as well as the plants' biomass, thus affecting ecosystem photosynthesis and transpiration rates and leading generally to a net increase in water use efficiency (Lemordant et al., 2018). These coupled changes have profound implications for the simulation of the carbon and water cycles (Gentine et al., 2019; see also Section 5.4.1), which can be better assessed with the new generation Earth system models, although both the carbon concentration and carbon-climate feedbacks remain highly uncertain over land (Arora et al., 2020) {5.4.5}. The water constraints on the terrestrial carbon sinks are a matter of debate regarding the feasibility or efficiency of

[1] In this chapter, we use evaporation to include all evaporative processes that include transpiration over land while the term evapotranspiration (ET) is also used interchangeably when the focus is only on land.

some land-based carbon dioxide removal and sequestration techniques requested to comply with the Paris Agreement (Fuss et al., 2018; Belyazid and Giuliana, 2019) {5.6.2.2.1}.

[START FIGURE 8.1 HERE]

Figure 8.1: Depiction of the water cycle based on previous assessments (Trenberth et al., 2011; Rodell et al., 2015; Abbott et al., 2019) with minor adjustments for groundwater flows (Kwon et al., 2014; Zhou et al., 2019c; Luijendijk et al., 2020), seasonal snow (Pulliainen et al., 2020) and ocean precipitation and evaporation (Stephens et al., 2012; Allan et al., 2020; Gutenstein et al., 2020). In the atmosphere, which accounts for only 0.001% of all water on Earth, water primarily occurs as a gas (water vapour), but it is also present as ice and liquid water within clouds. The ocean is the primary water reservoir on Earth, which is comprised mostly of liquid water across much of the globe, but also includes areas covered by ice in polar regions. Liquid freshwater on land forms surface water (lakes, rivers), soil moisture and groundwater stores, together accounting for 1.8% of global water (Stocker et al., 2013). Solid terrestrial water that occurs as ice sheets, glaciers, snow and ice on the surface and permafrost currently represents 2.2% of the planet's water (Stocker et al., 2013). Water that falls as snow in winter provides soil moisture and streamflow after melting, which are essential for human activities and ecosystem functioning.

[END FIGURE 8.1 HERE]

8.1.2 Summary of water cycle changes from AR5 and special reports

This Report is the first IPCC assessment to include a chapter specifically dedicated to providing an integrated assessment of the global water cycle changes, by building on many chapters from previous reports. This section summarises observed and projected water cycle changes reported in the AR5 (IPCC, 2013) and in the recent IPCC special reports on global warming of 1.5ºC (SR1.5), ocean and cryosphere in a changing climate (SROCC), and climate change and land (SRCCL).

8.1.2.1 Summary of observed and projected water cycle changes from AR5

Based on long-term observational evidence (Hartmann et al., 2013), AR5 concluded it was *likely* that anthropogenic influence has affected the water cycle since the 1960s (IPCC, 2018). Detectable human influence on changes to the water cycle were found in atmospheric moisture content (*medium confidence*), global-scale changes of precipitation over land (*medium confidence*), intensification of heavy precipitation events over land regions where sufficient data networks exist (*medium confidence*), and *very likely* changes to ocean salinity through its connection with evaporation minus precipitation change patterns (Stocker et al., 2013) {2.5, 2.6, 3.3, 7.6, 10.3, 10.4}. AR5 also reported that it is *very likely* that global surface air specific humidity increased since the 1970s. There was *low confidence* in the observations of global-scale cloud variability and trends, *medium confidence* in reductions of pan-evaporation, and *medium confidence* in the non-monotonic changes of global evapotranspiration since the 1980s. In terms of streamflow and runoff, the AR5 identified that there is *low confidence* in the observed increasing trends of global river discharge during the 20th century. Similarly, AR5 concluded that there is *low confidence* in any global-scale observed trend in drought or dryness (lack of rainfall) since the mid-20th century. Yet, the frequency and intensity of drought *likely* increased in the Mediterranean and West Africa, while they *likely* decreased in central North America and north-western Australia since 1950.

Water cycle projections in AR5 (Collins et al., 2013) were considered primarily in terms of water vapour, precipitation, surface evaporation, runoff, and snowpack. Globally-averaged precipitation was projected to increase with global warming with *virtual certainty* (12ES, 12.4.1.1). Regionally, precipitation in some areas of the tropics and polar regions could increase by more than 50% by the end of the 21st century under the RCP8.5 emissions scenario, while precipitation in large areas of the subtropics could decrease by 30% or more (AR5 FAQ 12.2, Figure 12.22). Overall, the contrast of annual mean precipitation between dry and wet regions and between dry and wet seasons ("wet-get-wetter and dry-get-drier") was projected to increase over most of the globe with *high confidence* (12ES, 12.4.5.2). Globally, the frequency of intense precipitation events was projected to increase while the frequency of all precipitation events was projected to decrease, leading to the contradictory-seeming projection of a simultaneous increase in both droughts and floods (AR5 FAQ 12.2, 12.4.5.5). Surface evaporation change was projected to be positive over most of the ocean and to generally follow the pattern of precipitation change over land (12ES, 12.4.5.4). Near-surface relative humidity reductions over many land areas were projected to be *likely*, with *medium confidence* (12.4.5.1). General decreases in soil moisture in present-day dry regions were considered *likely,* and projected with *medium confidence* under the RCP8.5 scenario (12.4.5.3). Soil moisture drying in the Mediterranean, southwest USA and southern African regions was considered *likely*, with *high confidence* by the end of this century under the RCP8.5 scenario (12.4.5.3). Projections for annual runoff included both decreases and increases. Decreases in Northern Hemisphere snow cover were assessed as *very likely* with continued global warming (12.4.6.2). As temperatures increase, snow accumulation was projected to begin later in the year and melting to start earlier, with related changes in snowmelt-driven river flows (AR5 FAQ 12.2, 12.4.6.2). In terms of the potential for abrupt change in components of the water cycle, long-term droughts and monsoonal circulation were identified as potentially undergoing rapid changes, but the assessment was reported with *low confidence* (Table 12.4, 12.5.5.8.1, 12.5.5.8.2).

8.1.2.2 Key findings of AR6 special reports

The SR1.5 assessed the impacts of global warming of 1.5°C above pre-industrial levels. The dominant human influence on observed global warming and related water cycle changes was confirmed. Further evidence that anthropogenic global warming has caused an increase in the frequency, intensity and/or amount of heavy precipitation events at the global scale (*medium confidence*), as well as in drought occurrence in the Mediterranean region (*medium confidence*) was also reported. Chapter 3 of SR1.5 (Hoegh-Guldberg et al., 2019) highlights that each half degree of additional global warming influences the climate response. Heavy precipitation shows a global tendency to increase more at 2°C compared to 1.5°C, though there is *low confidence* in projected regional differences in heavy precipitation at 1.5°C compared to 2°C global warming, except at high latitudes or at high altitude where there is *medium confidence*. A key finding is that "limiting global warming to 1.5°C compared to 2°C would approximately halve the proportion of the world population expected to suffer water scarcity, although there is considerable variability between regions (*medium confidence*)" (SR1.5). This is consistent with greater adverse impacts found at 2°C compared to 1.5°C for a number of dryness or drought indices (Schleussner et al., 2016; Lehner et al., 2017; Greve et al., 2018). There is also *medium confidence* that land areas with increased runoff and exposure to flood hazards will increase more at 2°C compared to 1.5°C of global warming.

The Special Report on the Ocean and Cryosphere in a changing Climate (SROCC) provides a comprehensive assessment of recent and projected changes, specifically in snow and ice-covered areas that form a key component of the water cycle in high-elevation and high-latitudes areas. High mountain regions have experienced significant warming since the early 20th century, resulting in reduced snowpack on average (Marty et al., 2017), with glaciers retreating globally since the mid-20th century (Marzeion 2018; Zemp et al., 2019). Glacier shrinkage and snow cover changes have led to changes (both increases and decreases) in streamflow in many mountain regions in recent decades (Milner et al., 2017). Permafrost regions have undergone degradation and ground-ice loss due to recent warming (Lu et al., 2017). Glacier mass loss is projected to continue through the 21st century under all scenarios. In high mountain areas, low-elevation snow cover is also projected to decrease, regardless of emissions scenario. Widespread permafrost thaw is projected to continue through this century and beyond. River runoff in snow- or glacier-fed basins is projected to increase in winter and to decrease in summer (and in the annual mean) by 2100. In the oceans, the Atlantic Meridional Overturning Circulation (AMOC) will *very likely* weaken over the 21st century under all emissions scenarios (SROCC), with potential effects on atmospheric circulation and the water cycle at the regional scale (cf. Section 8.6).

The Special Report on climate change, desertification, land degradation, sustainable management, food security, and greenhouse gas fluxes in terrestrial ecosystems (SRCCL) has clear connections with the water cycle. This report indicates that since 1850-1900, land surface temperature has risen nearly twice as much as global surface temperature (*high confidence*), with an increase in dry climates (*high confidence*). Land surface processes modulate the likelihood, intensity and duration of many extreme events including droughts (*medium confidence*) and heavy precipitation (*medium confidence*). The direction and magnitude of hydrological changes induced by land use change and land surface feedbacks vary with location and season (*high confidence*). Desertification exacerbates climate change through feedbacks involving vegetation cover, greenhouse gases and mineral dust aerosol (*high confidence*). Urbanisation increases extreme rainfall events over or downwind of cities (*medium confidence*). Intensification of rainy events increase their consequences on land degradation.

8.1.3 Chapter motivations, framing and preview

AR5 report was a major step forward in the assessment of the human influence on the Earth's water cycle, yet regional projections of precipitation and water resources often remained very uncertain for a range of reasons including modelling uncertainty and the large influence of internal variability (Deser et al., 2012; Hawkins and Sutton, 2011; see also Section 1.4.3 and 8.5.2). Since AR5, longer and more homogeneous observational and reanalysis datasets have been produced along with new ensembles of historical simulations driven by all or individual anthropogenic forcings. These factors, together with improved detection-attribution tools, has enabled a more comprehensive assessment and a better understanding of recent observed water cycle changes, including the competing effects of greenhouse gases and aerosol emissions. New paleoclimate reconstructions have been also developed, particularly from the Southern Hemisphere (SH), that were not available at the time of AR5. There have also been advances in modelling clouds, precipitation, surface fluxes, vegetation, snow, floodplains, ground water and other processes relevant to the water cycle. Convection permitting and cloud resolving models have been implemented over increasingly large domains and can be used as benchmarks for the evaluation of the current-generation climate models. The added value of increased resolution in global or regional climate models can be also assessed more thoroughly based on dedicated model intercomparison projects (see Section 10.3.3 and Section 8.5.1). Ongoing research activities on decadal predictions and observational constraints are aimed at narrowing the plausible range of near-term (2021-2040) to long-term (2081-2100) water cycle changes.

This chapter assesses water cycle changes and considers climate change from the perspective of its effects on water availability (including streamflow and soil moisture, snow mass and glaciers, groundwater, wetlands and lakes) rather than only precipitation. The chapter highlights the sensitivity of the water cycle to multiple drivers and the complexity of its responses, depending on regions, seasons and timescales. Anthropogenic drivers include not only emissions of greenhouse gases but also different species of aerosols, land and water management practices. Emphasis is placed on assessing the full range of projections, including 'low likelihood, high impact' climate trajectories such as the potential for abrupt changes in the water cycle.

The chapter starts with theoretical evidence that link small-scale processes and drivers, as well as global energy budget and large-scale circulation constraints to physically-understood changes in the global water cycle (Section 8.2). Observed and projected water cycle changes (Section 8.3 and 8.4, respectively) are assessed in separate sections, but with a parallel structure to facilitate comparison of a specific topic across sections. Projections are primarily assessed on the basis of contrasted emission scenarios to emphasize the water cycle response to mitigation. Unless otherwise specified, projected anomalies are estimated relative to the 1995-2014 baseline climatology and are assessed over 20-year timeslices, 2021-2040, 2041-2060 and 2081-2100 for near-, mid- and long-term changes respectively. Beyond multi-model ensemble means, model response uncertainty, the influence of natural climate variability, and the potential non-linearities in the regional water cycle response are also considered (Section 8.5). Low likelihood but physically plausible high-impact scenarios are also assessed, especially the potential for abrupt climate change (Section 8.6). Final remarks about future studies on water cycle changes (Section 8.7) are also provided, and the chapter addresses three frequently asked questions (FAQs) on the water cycle's sensitivity to land use change (FAQ 8.1), the projected occurrence and severity of floods (FAQ 8.2) and droughts (FAQ 8.3) at the global scale. This chapter outline is summarized with a schematic (Figure 8.2) which also provides a quick guide to the main topics addressed across the different sections.

[START FIGURE 8.2 HERE]

Figure 8.2: Schematic of the chapter structure and quick guide to the chapter content.

[END FIGURE 8.2 HERE]

Chapter 8 has multiple links across all AR6 WGI chapters, so necessarily includes references to other chapter subsections and figures. Model evaluation of large-scale circulation, precipitation, and hydrological extremes is mostly covered by Chapter 3 and 11, respectively, while Chapter 8 focuses on key processes relevant to the water cycle and their resolution-dependent representation in models. Observed and projected changes in large-scale circulation and precipitation are primarily assessed in Chapters 2, 3 and 4. Beyond global and regional mean precipitation amounts, Chapter 8 also focuses on other precipitation properties (e.g., frequency, intensity and seasonality) and other water cycle variables (evapotranspiration, runoff, soil moisture and aridity, solid and liquid freshwater reservoirs). Key regional phenomena (e.g., tropical overturning circulations, monsoons, extratropical stationary waves and stormtracks, modes of variability and related teleconnections) are also assessed given their major dynamical contribution to regional water cycle changes. Although the biosphere and the cryosphere are key components of the water cycle, a more comprehensive assessment of their responses can be found in Chapters 5 and 9, respectively. Further assessment on regional water cycle changes can be found in Chapters 10 to 12 and in the Atlas. The reader is also referred to the interactive Atlas for a more detailed assessment of the range of model biases and responses at the regional scale. Beyond WGI, water is also a major topic for both adaptation and mitigation policies so has strong connections with both WGII and WGIII. Assessment of hydrological impacts at basin and catchment scales, including a broader discussion on adaptation and vulnerability, potential threats to water security, societal responses, improving resilience in water systems and related case studies is provided in WGII (Chapter 4).

8.2 Why should we expect water cycle changes?

It is well understood that global precipitation and evaporation changes are determined by Earth's energy balance (Section 8.2.1). At regional scales smaller than ~4000 km, water cycle changes become dominated by the transport of moisture (Dagan et al., 2019a; Jakob et al., 2019; Dagan and Stier, 2020), which depend on both thermodynamic and dynamical processes (Section 8.2.2). The constraints of energy budgets at global scales and moisture budgets at regional scales cause key water cycle characteristics such as precipitation intensity, duration and intermittence to alter as the climate warms (Pendergrass and Hartmann, 2014b; Döll et al., 2018a). Future water availability is also determined by changes in evaporation, which is driven by a general increase in the atmospheric evaporative demand (Scheff and Frierson, 2014) and modulated by vegetation controls on evaporative losses (Milly and Dunne, 2016; Lemordant et al., 2018; Vicente-Serrano et al., 2020b). At regional scales, water cycle changes result from the interplay between multiple potential drivers (CO_2, aerosols, land use change and human water use; Section 8.2.3). This section assesses advances in physical understanding of global to regional drivers of water cycle changes.

8.2.1 Global water cycle constraints

The Clausius-Clapeyron equation determines that low-altitude specific humidity increases by about 7% per ^{o}C of warming, assuming that relative humidity remains constant, which is approximately true at a global scale but not necessarily valid regionally. It is *very likely* that near surface specific humidity has increased since the 1970s (Section 2.3.1) and total atmospheric water vapour content (precipitable water) is *very likely* to increase at close to a thermodynamic rate on average globally with continued warming. Different radiative forcing mechanisms lead to some variation in the global mean thermodynamic response by altering the relative humidity distribution: the rate of global precipitable water increase with global surface temperature ranges[2] from 6.4±1.5% per ºC for sulphate aerosol-induced changes to 9.8±3.3% per ºC for black carbon-induced changes based on idealised modelling (Hodnebrog et al., 2019b). Specific humidity increases at a lower rate over land due to decreasing relative humidity (Collins et al., 2013) as corroborated by observations and simple models (Byrne and O'Gorman, 2018). Prevalent increases in atmospheric water vapour drive powerful amplifying feedbacks (Section 7.4.2.2), intensify atmospheric moisture transport and heavy precipitation events (Section 8.2.3.2), and alter the surface and atmospheric energy balance, thereby influencing global evaporation and precipitation changes (Figure 8.3).

While thermodynamics exert a strong control on water vapour changes, global mean precipitation and evaporation are constrained by the balance of energy fluxes in the atmosphere and at the surface (Figure 8.3). Global mean precipitation increases of 1-3% per ºC of warming, as estimated in AR5 (Collins et al., 2013), are explained as a combination of rapid (or fast) atmospheric adjustments and slow temperature-driven responses (Figure 8.3, panels 1-4) to radiative forcings (Andrews et al., 2010; Bala et al., 2010; Cao et al., 2012). Fast atmospheric adjustments are caused by near-instantaneous (hours to days) changes in the atmospheric energy budget (Figure 8.3, panels 1-3) and atmospheric properties (e.g. temperature, clouds and water vapour) in direct response to the radiative effects of a forcing agent (Sherwood et al., 2015). A further relatively fast (days to months) adjustment of the climate system involves interactions with vegetation and land surface temperature (Figure 8.3, panel 3), which respond more rapidly than ocean temperature to a radiative forcing (Cao et al., 2012; Dong et al., 2014). The slower temperature-dependent precipitation response is driven by the increased atmospheric radiative cooling rate of a warming atmosphere. Warming drives increases in precipitation intensity while frequency is dominated by rapid atmospheric adjustments to the radiative forcing based on $4xCO_2$ CMIP6 simulations (Douville and John, 2020). Since AR5, many new studies applying the dual rapid adjustment and slow response framework show that global precipitation responses to different forcing agents are physically well understood (Fläschner et al., 2016; MacIntosh et al., 2016; Samset et al., 2016; Myhre et al., 2018a). Further confidence in the coupled processes involved are provided by simple models representing the energy budget and thermodynamic constraints that limit global mean evaporation to around 1.5% per °C (Siler et al., 2018b). This strengthens the physical link between energy budget and thermodynamic drivers of the global water cycle (Section 8.2.2.1).

[START FIGURE 8.3 HERE]

Figure 8.3: Schematic representation of fast and slow responses of the atmospheric energy balance and global precipitation to radiative forcing. The atmospheric energy budget ('baseline' panel) responds instantaneously to radiative forcings (1), leading to rapid atmospheric adjustments (2) and slower semi-rapid adjustments involving the land surface and vegetation that further modify atmospheric circulation patterns (3). This slow precipitation response to global mean surface air temperature (4) is quantified as the hydrological sensitivity, η, and the total precipitation response, including initial rapid adjustments, is termed the apparent hydrological sensitivity, η_a (a). The slow precipitation response over land and ocean develops over time (b–d). Large, filled arrows ('baseline'-4) depict fluxes or circulation change while small arrows (1-4) denote increases (↑) or decreases (↓) in variables (P is precipitation; L is atmospheric longwave radiative cooling, S is solar radiation absorption by the atmosphere; H is sensible heat flux; E is surface evaporative heat flux and T is temperature). (Adapted from Allan et al., 2020, Chapter 7 Figure 7.2 and Figure 8.1)

[END FIGURE 8.3 HERE]

Hydrological sensitivity (*η*) is defined as the linear change in global mean precipitation with global surface air temperature (GSAT) once rapid adjustments of the hydrological cycle to radiative forcings have occurred (Figure 8.3a). There is robust understanding and high agreement across idealised CO_2 forcing CMIP5 and CMIP6 experiments (Fläschner et al., 2016; Samset et al., 2018b; Pendergrass, 2020b) that η = 2.1-3.1 % per ºC (Figure 8.4). The magnitude of η depends primarily on atmospheric net radiative cooling which is controlled by thermal deepening of the troposphere (Jeevanjee and Romps, 2018) and limited by surface

[2] 5-95% confidence range estimates are quoted unless otherwise stated

evaporation and consequent atmospheric latent heat release and warming (Webb et al., 2018). Climate feedbacks (e.g. temperature lapse rate and clouds) that vary across models (Section 7.4; Section 3.8.2) also modulate the magnitude of η (O'Gorman et al., 2012; Fläschner et al., 2016; Richardson et al., 2018c). Uncertainty in η across CMIP5 models relating to deficiencies in representing low-altitude cloud feedbacks (Watanabe et al., 2018) and absorption of shortwave radiation by atmospheric water vapour (DeAngelis et al., 2015) do not apply well to CMIP6 simulations, the latter improvement explained by more accurate radiative transfer modelling (Pendergrass, 2020b).

Observed estimates of hydrological sensitivity (η = 3.2 ± 0.8 % per ºC) based on interannual variability (Allan et al., 2020) or responses to El Niño-Southern Oscillation (ENSO) of 9 % per ºC (Adler et al., 2017) are not suitable to assess the magnitude of η (Figure 8.4). This is because these relationships depend on amplifying feedbacks associated with ENSO-related cloud changes (Stephens et al., 2018b) that may not be relevant for longer term climate change. However, there is robust evidence and high agreement across observations, modelling and supporting physics that precipitation increases at a lower % per ºC rate than water vapour content in the global mean (Held and Soden, 2006b; Collins et al., 2013; Allan et al., 2020), implying an increased residence time of atmospheric water vapour (Hodnebrog et al., 2019b; Dijk et al., 2020). Increasing global precipitation, evaporation and moisture fluxes with warming thereby drive an intensification but not acceleration of the global water cycle (Sections 8.3.1.1 and 8.4.1.1).

The overall global mean rate of precipitation change per ºC of GSAT increase, **apparent hydrological sensitivity (η_a)**, is reduced compared to hydrological sensitivity by the direct influence of radiative forcing agents on the atmospheric energy balance. Rapid atmospheric adjustments that alter precipitation are primarily caused by GHGs and absorbing aerosols, with high agreement and medium evidence across idealised simulations (Fläschner et al., 2016; Samset et al., 2016). A range of rapid precipitation adjustments to CO_2 between models are also attributed to vegetation responses leading to a repartitioning of surface latent and sensible heat fluxes (DeAngelis et al., 2016). Values obtained from six CMIP5 models simulating the Last Glacial Maximum and pre-industrial period (η_a=1.6-3.0 % per ºC) are larger than for each corresponding 4xCO_2 experiment (η_a=1.3–2.6 % per ºC) due to differences in the mix of forcings, vegetation and land surface changes and a higher thermodynamic % per ºC evaporation scaling in the colder state (Li et al., 2013b). Updated estimates across comparable experiments from 22 CMIP5/CMIP6 models (Rehfeld et al., 2020) display a consistent range (η_a=1.7±0.6 % per ºC; Figure 8.4; Section 8.4.1.1). Confirming η_a in observations (Figure 8.4) is difficult due to measurement uncertainty, varying rapid adjustments to radiative forcing and unforced variability (Dai and Bloecker, 2019; Allan et al., 2020).

Climate drivers that instantaneously affect the surface much more than the atmospheric energy budget (such as solar forcing and sulphate aerosol) produce only a small rapid adjustment of the global water cycle and therefore larger η_a than drivers that immediately modulate the atmospheric energy budget such as GHGs and absorbing aerosol (Salzmann, 2016; Samset et al., 2016; Lin et al., 2018; Liu et al., 2018a). Thus, global precipitation appears more sensitive to radiative forcing from sulphate aerosols (2.8±0.7 % per ºC; $\eta_a \approx \eta$) than GHGs (1.4±0.5 % per ºC; $\eta_a < \eta$) while the response to black carbon aerosol can be negative (-3.5±5.0 % per ºC; $\eta_a << \eta$) due to strong atmospheric solar absorption (Samset et al., 2016). Therefore, artificially reducing surface absorbed sunlight through solar radiation modification strategies to mitigate GHG warming will not mitigate precipitation changes (see Sections 4.6.3.3; 6.4.7; 8.6.3). Aerosol-induced precipitation changes depend upon the type of aerosol species and their spatial distribution. Global mean precipitation increases after complete removal of present day anthropogenic aerosol emissions (see also Section 4.4.4) in four different climate models (η_a = 1.6-5.5% per ºC) are mainly attributed to sulphate aerosol as opposed to other aerosol species (Samset et al., 2018b). Idealised modelling studies show that sulphate aerosol increases over Europe produce a larger global precipitation response than an equivalent increase in aerosol burden or radiative forcing over Asia, explained by differences in cloud climatology and cloud-aerosol interaction (Kasoar et al., 2018; Liu et al., 2018c). The vertical profiles of black carbon and ozone further influence the magnitude of the rapid global precipitation response, yet are difficult to observe and simulate (Allen and Landuyt, 2014; MacIntosh et al., 2016; Stjern et al., 2017; Sand et al., 2020).

Hydrological sensitivity is generally lower over land but with a large uncertainty range (η = -0.1 to 3.0 % per ºC GSAT) relative to the oceans (η = 2.3 to 3.3 % per ºC) based on multi-model 4xCO_2 CMIP6 simulations

(Pendergrass, 2020b), broadly consistent with comparable CMIP5 experiments (Richardson et al., 2018c; Samset et al., 2018a). Suppressed hydrological sensitivity over land (Figure 8.3d; Figure 8.4) is associated with greater warming compared with the oceans, which alters atmospheric circulation and precipitation patterns (Saint-Lu et al., 2020). Also, since oceans supply much of the moisture to fuel precipitation over land, the slower ocean warming rate means there is insufficient moisture supplied to maintain continental relative humidity levels (Byrne and O'Gorman, 2018), which can inhibit convection (Chen et al., 2020b). Land surface feedbacks involving soil-vegetation-atmosphere coupling further drive continental drying (Berg et al., 2016; Kumar et al., 2016; Chandan and Peltier, 2020). The suppressed hydrological sensitivity is counteracted by rapid precipitation responses in most GHG-forced simulations, explained by greater surface downward longwave radiation due to CO_2 increases that rapidly warm the land, destabilize the troposphere and strengthen vertical motion in the short term (Chadwick et al., 2014; Richardson et al., 2016, 2018c). There is medium understanding of how land-sea warming contrast governs rapid precipitation responses based on idealised modelling that shows similar spatial patterns of precipitation response to radiative forcing from GHGs, solar forcing and absorbing aerosols (Xie et al., 2013; Samset et al., 2016; Kasoar et al., 2018). Rapid precipitation adjustments to CO_2 have been counteracted by cooling from anthropogenic aerosol increases over land (Box 8.1) but this compensation is expected to diminish as aerosol forcing declines (Richardson et al., 2018c). The fast and slow precipitation responses over global land globally combine during transient climate change (Figure 8.3d). This explains a consistent land and ocean mean precipitation increase in projections (Chapter 4, Table 4.3) but this is determined by a complex and model-dependent evolution of continental water cycle changes over space and time.

Increases in global precipitation over time, as the climate warms, are partly offset by the overall cooling effects of anthropogenic aerosol and by rapid atmospheric adjustments to increases in GHGs and absorbing aerosol. This explains why multi-decadal trends in global precipitation responses in the satellite era (Adler et al., 2017; Allan et al., 2020) are small and difficult to interpret given observational uncertainty, internal variability and volcanic forcings. The delayed warming effect of rising CO_2 concentration, combined with declining aerosol cooling, are expected to increase the importance of the slow temperature-related effects on the energy budget relative to the more rapid direct radiative forcing effects as transient climate change progresses (Shine et al., 2015; Salzmann, 2016; Myhre et al., 2018b).

In summary, there is *high confidence* that global mean evaporation and precipitation increase with global warming, but the estimated rate is model-dependent (*very likely* range of 2-3 % per ºC) The global increase in precipitation is determined by a robust response to global surface temperature only (*very likely* 2–3% per 1°C) that is partly offset by fast atmospheric adjustments to the vertical profile of atmospheric heating by GHGs and aerosols. Global precipitation increases due to GHGs are offset by the well-understood overall surface radiative cooling effect by aerosols (*high confidence*). Over land, the average warming-related increase in precipitation is expected to be smaller than over the ocean due to increasing land-ocean thermal contrast and surface feedbacks, but the overall precipitation increase over land is generally reinforced by fast atmospheric responses to GHGs that strengthens convergence of winds (*medium confidence*). Global mean precipitation and evaporation increase at a lower rate than atmospheric moisture per ºC of global warming (*high confidence*) leading to longer water vapour lifetime in the atmosphere and driving changes in precipitation intensity, duration and frequency and an overall intensification but not acceleration of the global water cycle.

8.2.2 Constraints on the regional water cycle

8.2.2.1 Thermodynamic constraints on atmospheric moisture fluxes

A warming climate drives increases in atmospheric moisture and horizontal moisture transport from the divergent to the convergent portions of the atmospheric circulation (including storm systems, the tropical rain belt and monsoons) that on average amplifies existing precipitation minus evaporation (P-E) patterns (Held and Soden, 2006a). Increased latent heat transports in high latitudes also contribute to polar amplification of warming (Section 7.4.4.1). Although convergent parts of the atmospheric circulation are expected to become wetter (in terms of increasing P-E) and net evaporative regions drier (increasing E-P) these regions are not geographically and seasonally fixed and their location and timing are expected to alter (Section 8.2.2.2). Atmospheric and ocean circulation changes overall decrease the amplification of P-E and salinity patterns. Paleoclimate evidence confirms that during the Last Glacial Maximum (21–19 thousand years ago), zonal mean changes were roughly in agreement with thermodynamic expectations (Li et al., 2013b). However regional changes can be dominated by dynamics, including responses to the large Northern Hemisphere ice sheets (DiNezio and Tierney, 2013; Bhattacharya et al., 2017b; Scheff et al., 2017; D'Agostino et al., 2019; Lowry and Morrill, 2019) such that altered P-E patterns are not well described by thermodynamic drivers (Oster et al., 2015; Lora, 2018; Morrill et al., 2018).

There is robust evidence and high agreement across thermodynamics, detailed modelling and observations that amplification of P-E patterns occurs over the oceans (Figure 8.5a) with an associated "fresh gets fresher, salty gets saltier" signature in ocean salinity (Sections 2.3.3.2 and 3.5.2). This amplification is moderated by proportionally larger increases in sub-tropical ocean evaporation and weakening of the tropical circulation (Section 8.2.2.2), an expectation supported by observations (Skliris et al., 2016) and process understanding (Yang and Roderick, 2019). Thermodynamics explain a smaller low latitude evaporation increase (1% per °C) than in high latitudes (5% per °C) with changes in surface radiation, boundary layer adjustments and ocean heat uptake playing a secondary role, based on idealised modelling (Siler et al., 2018b). Increased evaporation from warmer oceans and lakes is exacerbated by the loss of surface ice in some regions (Bintanja and Selten, 2014; Laîné et al., 2014; Wang et al., 2018d; Sharma et al., 2019; Woolway et al., 2020). This can generate a more local moisture source for precipitation, for example in northwest Greenland during non-summer months since the 1980s (Nusbaumer et al., 2019), though moisture transport changes can counteract this effect (Nygård et al., 2020). Ocean stratification due to heating of the upper layers through radiative forcing has been identified as a mechanism that further amplifies surface salinity patterns beyond the responses driven by water cycle changes alone (Zika et al., 2018).

[START FIGURE 8.4 HERE]

Figure 8.4: Estimate (5-95% range) of the increase in precipitation and its extremes with global mean surface warming. Global time averaged precipitation changes (left) are based on responses to increasing CO_2 (apparent hydrological sensitivity, η_a) and the temperature-dependent component (hydrological sensitivity, η) based on GCM experiments and including the land (L) and ocean (O) components (Fläschner et al., 2016; Richardson et al., 2018c; Samset et al., 2018a; Pendergrass, 2020b; Rehfeld et al., 2020) and observational estimates (GPCP/HadCRUTv4.6) using trends (1988-2014) as a proxy for η_a and interannual variability as a proxy for η with 90% confidence range accounting for statistical uncertainty only (Adler et al., 2017; Allan et al., 2020). For extreme precipitation, assessment is for 24 hour 99.9th percentile or annual maximum extremes from GCMs (Fischer and Knutti, 2015; Pendergrass et al., 2015; Borodina et al., 2017; Pfahl et al., 2017; Sillmann et al., 2017), regional climate models (RCMs) (Bao et al., 2017), an observationally constrained tropical estimate (O'Gorman, 2012) and estimates from observed changes (Westra et al., 2013; Donat et al., 2016; Borodina et al., 2017; Sun et al., 2020; Zeder and Fischer, 2020). For hourly and sub-hourly extremes observed changes (Barbero et al., 2017; Guerreiro et al., 2018) and high resolution models including RCM and cloud resolving models (CRMs) are assessed (Ban et al., 2015; Prein et al., 2017; Haerter and Schlemmer, 2018; Hodnebrog et al., 2019a; Lenderink et al., 2019). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.4 HERE]

Since AR5, numerous studies have confirmed that changes in P-E with warming over land cannot be interpreted simply as a "wet regions get wetter, dry regions gets drier" response (Chadwick et al., 2013; Greve et al., 2014; Roderick et al., 2014; Byrne and O'Gorman, 2015; Scheff and Frierson, 2015). Firstly, P-E is a simplistic diagnostic of the water cycle that inadequately describes "dryness" or aridity (Fu and Feng, 2014; Roderick et al., 2014; Greve and Seneviratne, 2015; Scheff and Frierson, 2015; Greve et al., 2019; Vicente-Serrano et al., 2020b). Secondly, terrestrial P-E is generally positive and balanced by surface runoff and percolation into subsurface soils and aquifers (Figure 8.1). As a result, the simple thermodynamic scaling (Figure 8.5b) predicts that P-E over land will become more positive (wetter) with warming (Greve et al., 2014; Roderick et al., 2014; Byrne and O'Gorman, 2015). This is not necessarily true, however, in the dry seasons and regions where terrestrial water is lost to the atmosphere and exported (Sheffield et al., 2013; Kumar et al., 2015; Keune and Miralles, 2019). Thirdly, regional P-E patterns over land are affected by changes in atmospheric circulation, oceanic moisture supply and land surface feedbacks. As the land warms more than oceans, spatial gradients in temperature and relative humidity influence moisture supply and reduce P-E over some land regions, such as southern Chile and Argentina around 30-50ºS as captured by an extended thermodynamic scaling (Figure 8.5b). Drying of soils can be amplified by vegetation responses (Berg et al., 2016; Byrne and O'Gorman, 2016; Lambert et al., 2017) but limited by atmospheric circulation feedbacks (Zhou et al., 2021). Changes in soil moisture and rainfall intensity (Sections 8.2.3.2-8.2.3.3) can alter the partitioning of precipitation between evaporation and runoff, further complicating terrestrial P-E responses (Short Gianotti et al., 2020).

The strong physical basis for regionally and seasonally dependent responses of P-E and the expectation for an increasing contrast between wet and dry seasons and weather regimes is supported by high agreement across multiple observational and CMIP5/CMIP6 modelling studies (Liu and Allan, 2013; Kumar et al., 2015; Polson and Hegerl, 2017; Ficklin et al., 2019; Deng et al., 2020; Schurer et al., 2020). Increased moisture transports into storm systems, monsoons and high latitudes increase the intensity of wet events (Section 8.2.3.2), while stronger atmospheric evaporative demand with warming (Scheff and Frierson, 2014; Vicente-Serrano et al., 2018; Cook et al., 2019) is an important mechanism for intensifying dry events (Section 8.2.3.3) and decreasing soil moisture over many subtropical land regions. However, aridification is modulated regionally by poleward migration of the sub-tropical dry zones and an increasing land-ocean temperature contrast that drives declining relative humidity (Section 8.2.2.2).

To summarise, increased moisture transport from evaporative oceans to high precipitation regions of the atmospheric circulation will drive amplified P-E and salinity patterns over the ocean (*high confidence*) while more complex regional changes are expected over land. Greater warming over land than ocean alters atmospheric circulation patterns and on average reduces continental near-surface relative humidity which along with vegetation feedbacks can contribute to regional decreases in precipitation (*high confidence*). Based on an improved understanding of thermodynamic drivers since AR5 and multiple lines of evidence, there is *high confidence* that very wet or dry seasons and weather patterns will intensify in a warming climate such that wet spells become wetter and dry spells drier.

8.2.2.2 Large-scale responses in atmospheric circulation patterns

Responses of the large-scale atmospheric circulation to a warming climate are not as well understood as thermodynamic drivers (Shepherd, 2014). AR5 identified robust features including a weakening and broadening of tropical circulation with poleward movement of tropical dry zones and mid-latitude jets (Collins et al., 2013). These can dominate regional water cycle changes, affecting the availability of fresh water and the occurrence of climate extremes. Atmospheric circulation changes generally dominate the spatial pattern of rapid precipitation adjustments (Section 8.2.1) to different forcing agents in the tropics (Bony et al., 2013; He and Soden, 2015; Richardson et al., 2016, 2018c; Tian et al., 2017; Li et al., 2018b). Radiative forcing with heterogeneous spatial patterns such as ozone and aerosols (including cloud interactions; Box 8.1; Section 6.4.1) drive substantial responses in regional atmospheric circulation through uneven heating and cooling effects (Liu et al., 2018c; Wilcox et al., 2018b; Dagan et al., 2019b). Changes in atmospheric circulation are also driven by slower, evolving patterns of warming and associated changes in temperature and moisture gradients (Bony et al., 2013; Samset et al., 2016, 2018a; Ceppi et al., 2018; Ma et al., 2018). There is strong evidence that large regional water cycle changes arise from the atmospheric circulation response to radiative forcings and associated SST pattern evolution but low agreement in the sign and magnitude (Chadwick et al., 2016a). The role of prolonged weather regimes in determining wet and dry extremes is also better understood since AR5 (Kingston and McMecking, 2015; Schubert et al., 2016; Richardson et al., 2018a; Barlow et al., 2019). Advances in knowledge of expected large-scale dynamical responses of the water cycle are further assessed in this section (see also Figure 8.21).

Long-term weakening of the tropical atmospheric overturning circulation is expected as climate warms in response to elevated CO_2 (Collins et al., 2013). A weaker circulation is required to reconcile global mean low-level water vapour increases (around 7% per ºC) with the smaller global precipitation responses of about 1-3% per ºC (Section 8.2.1). The slowdown can occur in both the Hadley and Walker circulations, but occurs preferentially in the Walker circulation in most climate models (Vecchi and Soden, 2007) but this response has been questioned on the basis of model bias in east Pacific SST (Seager et al., 2019a). Weakening is expected to drive P-E decreases over the western Pacific and increases over the eastern Pacific. However, the driving mechanisms for Walker circulation weakening differ to those involved in determining ENSO variability, so it is too simplistic to interpret changes as an El Niño pattern of regional hydrological cycle extremes (Sohn et al., 2019). Internal variability is also capable of temporarily strengthening the Walker circulation (Section 2.3.1.4.1)(L'Heureux et al., 2013; Chung et al., 2019) while regional responses depend on the pattern of warming (Sandeep et al., 2014).

Model simulations show a stronger Pacific Walker circulation during the Last Glacial Maximum in response to a cooler climate (consistent with an expected weakening in a warmer climate), but a weaker Indian Ocean east-west circulation in response to the exposure of the Sunda and Sahul shelves due to lowered sea level (DiNezio et al., 2011). The latter effect is detectable in proxies for hydroclimate, as well as salinity and sea-surface temperature (DiNezio and Tierney, 2013; DiNezio et al., 2018). More relevant to future warming is the mid-Pliocene period (3 million years ago), the last time the Earth experienced CO_2 levels comparable to present (see Cross-Chapter Box 2.4). Sea surface temperature (SST) reconstructions show a weakening of the Pacific zonal gradient and a pattern of warmth consistent with a weaker Walker cycle response (Corvec and Fletcher, 2017; Tierney et al., 2019; McClymont et al., 2020). Although the Pliocene SST pattern and wet subtropics contrast with present conditions (Burls and Fedorov, 2017), the paleoclimate record strengthens evidence that a warmer climate is associated with a weaker Walker circulation (Cross-Chapter Box 2.4; Section 3.3.3).

Since AR5, weakening of the tropical circulation has been explained as a rapid response to increasing CO_2 concentrations and slower response to warming and evolving SST patterns (He and Soden, 2017; Xia and Huang, 2017; Shaw and Tan, 2018; Chemke and Polvani, 2020). Large-scale tropical circulation weakens by 3-4% in a rapid response to a quadrupling of CO_2 concentrations (Plesca et al., 2018), which suppresses tropospheric radiative cooling, particularly in sub-tropical ocean subsidence regions (Bony et al., 2013; Merlis, 2015; Richardson et al., 2016). The resulting increased atmospheric stability explains the rapid weakening of the Walker circulation (Wills et al., 2017) and Northern Hemisphere Hadley Cell (Chemke and Polvani, 2020). Subsequent surface warming contributes up to a 12% slowing of circulation for a uniform 4ºC SST increase, driven by thermodynamic decreases in temperature lapse rate (Plesca et al., 2018).

The regional Intertropical Convergence Zone (ITCZ) position, width and strength determine the location and seasonality of the tropical rain belt. Since AR5, multiple studies have linked cross-equatorial energy transport to the mean ITCZ position (Donohoe et al., 2013; Frierson et al., 2013; Bischoff and Schneider, 2014; Boos and Korty, 2016; Loeb et al., 2016; Adam et al., 2018; Biasutti and Voigt, 2019). Multi-model studies agree that aerosol cooling in the Northern Hemisphere led to a southward shift in the ITCZ and tropical precipitation after the 1950s up to the 1980s that is linked with the 1980s Sahel drought (Box 8.1; Section 8.3.2.4; Section 10.4.2.1). In particular, aerosol-cloud interaction was identified as a potentially important driver of this shift (Chung and Soden, 2017) but this is uncertain since observations suggest that models may overestimate (Malavelle et al., 2017; Toll et al., 2017) or underestimate (Rosenfeld et al., 2019) the aerosol cloud-mediated cooling effects. In addition, greenhouse gas forcing has been invoked in explaining much of the increase in Sahel precipitation since the 1980s through enhanced meridional temperature gradient, with only a secondary role for aerosol (Dong and Sutton, 2015).

Understanding of how ITCZ width and strength respond to a warming climate has improved since AR5 (Byrne and Schneider, 2016; Harrop and Hartmann, 2016; Popp and Silvers, 2017; Dixit et al., 2018; Zhou et al., 2020). Studies suggest that convection gets stronger and more focused within the core of the ITCZ (Lau and Kim, 2015; Byrne et al., 2018). This leads to drying on the equatorward edges of the ITCZ and a moistening tendency in the ITCZ core (Byrne and Schneider, 2016). Feedbacks involving clouds have been identified as an important mechanism leading to tightening and strengthening of the ITCZ (Popp and Silvers, 2017; Su et al., 2017, 2019, 2020; Talib et al., 2018). Stronger ascent in the core amplifies the "wet get wetter" response while reduced moisture inflow near the ITCZ edges reduces this response below the 7% per ^{o}C thermodynamic increase in moisture transport. Thus, there is a range of evidence and medium agreement for strengthening and contraction of the ITCZ with warming that sharpens contrasts between wet and dry regimes. However, understanding of how the regional ITCZ location responds in a warming climate is not robust (Section 8.4.2.1) with *limited evidence* of distinct regional responses to GHG forcing including a northward shift over eastern Africa and the Indian Ocean and a southward shift in the eastern Pacific and Atlantic oceans (Mamalakis et al., 2021). Paleoclimate evidence highlights the distinct regional ITCZ responses to hemispheric asymmetry in volcanic and orbital forcing (McGee et al., 2014; Boos and Korty, 2016; Colose et al., 2016; Denniston et al., 2016; PAGES Hydro2K Consortium, 2017; Singarayer et al., 2017; Atwood et al., 2020) and rapid (>1^{o} latitude over decades) shifts in the ITCZ and regional monsoons in response to AMOC collapse cannot be ruled out (Sections 8.6.1.1 and 5.1.3).

Monsoons are key components of the tropical overturning circulation that can be understood as a balance between net energy input (e.g. radiative and turbulent fluxes) and the export of moist static energy. This 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 (Jalihal et al., 2019). Thermodynamic increases in moisture transport are expected to increase monsoon strength and area (Christensen et al., 2013). Since AR5, evidence continues to demonstrate that monsoon circulation is sensitive to spatially varying radiative forcing by anthropogenic aerosols (Hwang et al., 2013; Allen et al., 2015b; Li et al., 2016c) and GHGs (Dong and Sutton, 2015). Changes in SST patterns also play a role (Guo et al., 2016; Zhou et al., 2019b; Cao et al., 2020) by altering cross-equatorial energy transports and land-ocean temperature contrasts. This evidence continues to support a thermodynamic strengthening of monsoon precipitation that is partly offset by slowing of the tropical circulation but with *weak evidence* and *low agreement* for regional aspects of circulation changes. Disagreement between paleo-climate and modern observations, physical theory and numerical simulations of global monsoons have been partly reconciled (Section 3.3.3.2) through improved understanding of regional processes (Harrison et al., 2015; Bhattacharya et al., 2017a, 2018; Biasutti et al., 2018; D'Agostino et al., 2019; Jalihal et al., 2019; Seth et al., 2019), although interpreting past changes in the context of future projections requires careful account of differing forcings and feedbacks (D'Agostino et al., 2019). Assessment of past changes and future projections in regional monsoons are provided in Sections 2.3.1.4.2, 8.3.2.4 and 8.4.2.4.

Since AR5, understanding of poleward expansion of the Hadley Cells has improved (Section 2.3.1.4.1) but its role in subtropical drying is limited to the zonal mean and dominated by ocean regions (Byrne and O'Gorman, 2015; Grise and Polvani, 2016; He and Soden, 2017; Schmidt and Grise, 2017; Siler et al., 2018a; Chemke and Polvani, 2019; Grise and Davis, 2020). Over subtropical land, evolving SST patterns and land-ocean warming contrasts, that are partly explained by rapid responses to CO_2 increases, can dominate aspects of the atmospheric circulation response (Byrne and O'Gorman, 2015; He and Soden, 2015; Chadwick et al., 2017; Yang et al., 2020a) and resultant regional water cycle changes, particularly for projected drying in semi-arid, winter-rainfall dominated sub-tropical climates (Deitch et al., 2017; Brogli et al., 2019; Seager et al., 2019b; Zappa et al., 2020). Poleward expansion of the tropical belt is expected to drive a corresponding shift in mid-latitude storm tracks, but the controlling mechanisms differ between hemispheres. Southern Hemisphere expansion is driven by GHG forcing and amplified by stratospheric ozone depletion, while weaker Northern Hemisphere expansion in response to GHG forcing is modulated by tropospheric ozone and aerosol forcing, particularly black carbon (Davis et al., 2016; Grise et al., 2019; Watt-Meyer et al., 2019; Zhao et al., 2020). However, internal variability is found to dominate observed responses in the Northern Hemisphere, precluding attribution to radiative forcing (D'Agostino et al., 2020b). Paleoclimate evidence of poleward expansion and weakening of westerly winds in both hemispheres in the warmer Pliocene is linked to reduced equator to pole thermal gradients and ice volume (Abell et al., 2021).

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 (Cross Chapter Box 10.1). The thermal gradient between polar and lower latitude regions decreases at low levels due to Arctic warming amplification. However, at higher altitudes, the corresponding thermal gradient increases with warming due to cooling of the Arctic stratosphere and this is consistent with a strengthening of the winter jet stream in both hemispheres, yet there is *low agreement* on the precise mechanisms (Vallis et al., 2015; Vihma et al., 2016). Changes in the strength of the polar stratospheric vortex can also alter the mid-latitude circulation in winter, but responses are not consistent across models (Oudar et al., 2020a). Nevertheless, thermodynamic strengthening of moisture convergence into weather systems and polar regions is robust (Section 8.2.2.1) and remains valid despite weak understanding of atmospheric circulation change.

In summary, there is *high confidence* that altered atmospheric wind patterns in response to radiative forcing and evolving surface temperature patterns will affect the regional water cycle in most regions. Mean tropical circulation is expected to slow with global warming (*high confidence*) but temporary multi-decadal strengthening is possible due to internal variability (*medium confidence*). Slowing of the tropical circulation reduces the meridional P-E gradient over the Pacific and can partly offset thermodynamic amplification of P-E patterns and strengthening of monsoons (*high confidence*) but regional characteristics of tropical rain belt changes are not well understood. There is *medium confidence* in processes driving strengthening and tightening of the ITCZ that increase the contrasts between wet and dry tropical weather regimes and seasons. There is *high confidence* in understanding of how radiative forcing and global warming drive a poleward expansion of the subtropics and mid-latitude stormtracks but only *low confidence* in how poleward expansion influences drying of sub-tropical and mid-latitude climates. There is *low confidence* in understanding how Arctic warming amplification affects mid-latitude regional water cycles but *high confidence* that thermodynamic strengthening of precipitation within weather systems and in monsoons and polar regions is robust to large-scale circulation changes.

8.2.3 Local-scale physical processes affecting the water cycle

Processes operating at local scales are capable of substantially modifying the regional water cycle. This section assesses the development in understanding of processes affecting the atmosphere, surface and subsurface, including cryosphere and biosphere interactions and the direct impacts of human activities.

8.2.3.1 Hydrological processes related to ice and snow

Declining ice sheet mass, glacier extent and Northern Hemisphere sea ice, snow cover and permafrost (Collins et al., 2013; Vaughan et al., 2013) is an expected consequence of a warming climate (Sections 2.3.2; 3.4; 4.3.2.1; 9.3-9.5). A decline in mountain snow cover and increased snow and glacier melt will alter the amount and timing of seasonal runoff in mountain regions (Sections 3.4.2; 3.4.3; 9.5). Earlier and more extensive winter and spring snowmelt (Zeng et al., 2018a) can reduce summer and autumn runoff in snow dominated river basins of mid-high latitudes of the Northern Hemisphere (Rhoades et al., 2018; Blöschl et al., 2019). Since AR5, an earlier but less rapid snowmelt has been explained by reduced winter snowfall and less intense solar radiation earlier in the season (Musselman et al., 2017; Wu et al., 2018b; Grogan et al., 2020). Reduced snow cover also increases energy available for evaporation, which can dominate declining river discharge based on modelling of the Colorado River (Milly and Dunne, 2020). An increase in the fraction of precipitation falling as rain versus snow can lead to declines in both streamflow and groundwater storage in regions where snow melt is the primary source of recharge (Earman and Dettinger, 2011; Berghuijs et al., 2014). Such regions include western South America and western North America, semi-arid regions which rely on snowmelt from high mountain chains (Ragettli et al., 2016; Milly and Dunne, 2020). Rain-on snow melt events reduce at lower altitudes due to declining snow cover but increased at higher altitudes where snow tends to be replaced by rain based on observations and modelling (Musselman et al., 2018; Pall et al., 2019), thereby altering seasonal and regional characteristics of flooding (Section 11.5).

Seasonal meltwater from high mountain glaciers in Asia (see Cross-Chapter Box 10.4) supply the basic needs of 221±97 million people (Pritchard, 2019; Immerzeel et al., 2020). Glacier-melt in response to warming can initially lead to increased runoff volumes, especially in peak summer flows, but they will eventually decline as most glaciers continue to shrink. SROCC concluded there is *high confidence* that the peak runoff has already been passed for some smaller glaciers (Hock et al., 2019b). Increased precipitation and glacier melt can also contribute to rising lake levels and flood hazards in regions such as the inner

Tibetan Plateau, Patagonia, Peru, Alaska and Greenland (Lei et al., 2017; Shugar et al., 2020; Stuart-Smith et al., 2020). Since AR5, evidence from multiple locations (New Zealand, Greenland, Antarctica) shows that intrusions of warm, moist air are important in controlling glacier mass balance, the likelihood of extreme ablation or snowfall events depending on air temperature (Gorodetskaya et al., 2014; Mackintosh et al., 2017; Mattingly et al., 2018; Oltmanns et al., 2018; Little et al., 2019; Wille et al., 2019; Adusumilli et al., 2021). Sensible heating from warm air and increased longwave radiation from atmospheric moisture and low clouds drive melt events (Stuecker et al., 2018).

Reductions in snow, freshwater ice and permafrost affect terrestrial hydrology. Permafrost degradation reduces soil ice and alters the extent of thermokarst lake coverage (Section 9.5.2; SROCC (Meredith et al., 2019b)). A lag between current climate change and permafrost degradation is expected, given the slow response rates in frozen ground and the fact that snow cover insulates soil from sensible heat exchanges with the air above (Hoegh-Guldberg et al., 2018; García-García et al., 2019; Soong et al., 2020). Post-wildfire areas are also linked with permafrost degradation in the Arctic based on satellite observations (Yanagiya and Furuya, 2020). An increase in spring rainfall can increase heat advection by infiltration, exacerbating permafrost thaw and leading to increased methane emissions (Neumann et al., 2019) (Section 5.4.7). Increased heat transport by Arctic rivers can also contribute to earlier sea ice melt (Park et al., 2020).

In summary, it is *virtually certain* that warming will cause a loss of frozen water stores, except in areas where temperatures remain below 0ºC for most of the year. There is *high confidence* that warming and reduced snow volume drives an earlier snowmelt, leading to seasonally dependent changes in streamflow. There is *medium confidence* that weaker sunlight earlier in the season can reduce the rate of snowmelt. Melting of snowpack or glaciers can increase stream flow in high latitude and high-altitude catchments until frozen water reserves are depleted (*high confidence*). There is *high confidence* that warm, moist airflows and associated precipitation dominate glacier mass balance in some regions (New Zealand, Greenland, Antarctica).

8.2.3.2 Processes determining heavy precipitation and flooding

Evidence that heavy precipitation events (from sub-daily up to seasonal timescales) intensify as the planet warms has strengthened since AR5 (Box 11.1; Section 11.4; Cross Chapter Box 3.2) based on improved physical understanding, extensive modelling and increasing observational corroboration (O'Gorman, 2015; Fischer and Knutti, 2016; Neelin et al., 2017). There is *robust evidence*, with *medium agreement* across a range of modelling and observational studies, of thermodynamic intensification of wet seasons (Chou et al., 2013; Liu and Allan, 2013; Dunning et al., 2018; Lan et al., 2019; Zhang and Fueglistaler, 2019). Extreme daily precipitation is expected to increase at close to the 7%/ºC increase in the near-surface atmospheric moisture holding capacity determined by the Clausius-Clapeyron equation (Section 11.4, Figure 8.4), with *limited evidence* that higher rates apply for shorter duration precipitation events (Formayer and Fritz, 2017; Lenderink et al., 2017; Ali et al., 2018; Guerreiro et al., 2018; Burdanowitz et al., 2019; Zhang et al., 2019b). However, observed estimates sample multiple synoptic weather states, mixing thermodynamic and dynamic factors, so are not directly relatable to climate change responses (Bao et al., 2017; Drobinski et al., 2018). The contrasting spatial scales sampled by the observations and models (from global to cloud resolving) explain the large range of daily and sub-daily precipitation scaling with temperature assessed in Figure 8.4.

[START FIGURE 8.5 HERE]

Figure 8.5: Zonally-averaged annual mean changes in precipitation minus evaporation (P-E) over (a) ocean and (b) land between the historical (1995–2014) and SSP2-4.5 (2081–2100) CMIP6 simulations (blue lines, an average of the CanESM5 and MRI-ESM2-0 models). Dashed lines show estimated P-E changes using a simple thermodynamic scaling (Held and Soden, 2006b); dotted lines show estimates using an extended scaling (Byrne and O'Gorman, 2016). All curves have been smoothed in latitude using a three grid-point moving-average filter. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.5 HERE]

Since AR5, advances in understanding the expected changes in intense rainfall at the sub-daily time-scale (Section 11.4, Figure 8.4) are provided by idealised or high resolution model experiments and observations (Westra et al., 2014; Fowler et al., 2021). There is *robust evidence* from simplified calculations, convection resolving models and observations that thermodynamics drives an increase in convective available potential energy (CAPE) with warming and therefore the intensity of convective storms (Singh and O'Gorman, 2013; Romps, 2016; Barbero et al., 2019). Also, declining relative humidity over land (Sections 2.3.1.3.2; 8.2.2.1) increases lifting condensation level, thereby delaying but intensifying convective systems (Louf et al., 2019; Chen et al., 2020b). Larger systems are linked with increasing tropopause height (Lenderink et al., 2017) that can also amplify storm precipitation (Prein et al., 2017). However, the heaviest rainfall is not necessarily associated with the most intense (deepest) storms based on satellite data (Hamada et al., 2015; Hamada and Takayabu, 2018). Precipitation intensification can exceed thermodynamic expectations where and when additional latent heating invigorates individual storms (Section 11.4.1) as implied by *medium agreement* across modelling and observational studies (Berg et al., 2013; Molnar et al., 2015; Scoccimarro et al., 2015; Prein et al., 2017; Zhou and Wang, 2017; Nie et al., 2018; Zhang et al., 2018f; Kendon et al., 2019). This intensification depends on time of day, based on convection-permitting simulations (Meredith et al., 2019a).

Intensification of sub-daily rainfall is inhibited in regions and seasons where available moisture is limited (Prein et al., 2017). However, a fixed threshold temperature above which precipitation is limited by moisture availability is not supported by modelling evidence (Neelin et al., 2017; Prein et al., 2017). Enhanced latent heating within storms can also suppress convection at larger-scales due to atmospheric stabilization as demonstrated with high resolution, idealised and large ensemble modelling studies (Loriaux et al., 2017; Chan et al., 2018; Nie et al., 2018; Tandon et al., 2018; Kendon et al., 2019). Stability is also increased by the direct radiative heating effect of higher CO_2 concentrations (Baker et al., 2018) and influenced by aerosol effects on the atmospheric energy budget and cloud development (Box 8.1). Since AR5, modelling evidence shows increases in convective precipitation extremes are limited by droplet/ice fall speeds (Singh and O'Gorman, 2014; Sandvik et al., 2018) but these processes are only crudely represented (Tapiador et al., 2019b). Idealised regional and coupled global models combined with *limited* observational *evidence* shows that instantaneous precipitation extremes are sensitive to microphysical processes while daily extremes are determined more by the degree of convective aggregation (Bao and Sherwood, 2019; Pendergrass, 2020a).

Dynamical changes modify and can dominate thermodynamic drivers of local rainfall and flood hazard change (Box 11.1). For example, increased land-ocean temperature gradients (Section 8.2.2.2) explain more intense rain from convective systems over the Sahel based on satellite data since the 1980s (Taylor et al., 2017) and dynamical feedbacks can invigorate active to break phase transition over India (Karmakar et al., 2017; Roxy et al., 2017). Satellite data shows long-lived, organised mesoscale convective systems contribute disproportionally to extreme tropical precipitation (Roca and Fiolleau, 2020). Since AR5, the spatial variability in soil moisture has been linked with the timing and location of convective rainfall by altering the partitioning between latent and sensible heating. This was demonstrated for the Sahel, Europe and India in observations (Taylor et al., 2013a; Taylor, 2015; Petrova et al., 2018; Barton et al., 2019; Klein and Taylor, 2020) but depends on the moisture convergence regime (Welty et al., 2020). Only high resolution convection permitting models can capture the sub-grid scale mechanisms for convective initiation (Taylor et al., 2013a; Moon et al., 2019a). There is *medium evidence* that greater tropical cyclone rainfall totals can be caused by dynamical feedbacks (Chauvin et al., 2017) and slower propagation speed as tropical circulation weakens (Kossin, 2018). These processes amplify the thermodynamic intensification of rainfall (Section 11.7.1.2), yet observational support is weak (Chan, 2019; Lanzante, 2019; Moon et al., 2019b; Knutson et al., 2020). Slower decay following landfall, explained by larger stores of heat and moisture at higher SSTs, can also amplify rainfall amount based on observations and modelling (Li and Chakraborty, 2020). Rainfall intensity from the outer rain bands of tropical cyclones is also increased by aerosol-cloud interactions (Box 8.1).

The amount and intensity of rainfall within extratropical storms is expected to increase with atmospheric moisture. This is particularly evident for atmospheric rivers (see glossary) and research since AR5 has confirmed their link with flooding and terrestrial water storage (Froidevaux and Martius, 2016; Paltan et al., 2017; Waliser and Guan, 2017; Adusumilli et al., 2019; Ionita et al., 2020; Payne et al., 2020). There is *robust evidence* based on simple physics and detailed modelling that extra-tropical cyclone rainfall, including atmospheric river events, will intensify through increased atmospheric moisture flux (Lavers et al., 2013; Ramos et al., 2016; Yettella and Kay, 2017; Espinoza et al., 2018b; Algarra et al., 2020; Xu et al., 2020; Zavadoff and Kirtman, 2020; Zhao, 2020), although changes in dynamical aspects will modify responses regionally (section 8.4.2.8). For example, stronger latitudinal temperature gradients in the high latitude upper troposphere drive increased extra-tropical storm speed around 30-70ºN based on CMIP5 simulations (Dwyer and O'Gorman, 2017), causing reduced precipitation accumulation.

The response of flood hazard to changing rainfall characteristics depends on time and space scale and the nature of the land surface (Section 11.5.1; FAQ 8.2). Sustained and heavy rainfall can lead to widespread flooding and landslides while intensification of short-duration intense rainfall can increase the severity and frequency of flash flooding (Marengo et al., 2013; Chan et al., 2016; Gariano and Guzzetti, 2016; Sandvik et al., 2018). Flooding events in many tropical regions (e.g. north west South America, southern Africa and Australasia) are associated with ENSO variability (Emerton et al., 2017; Takahashi and Martínez, 2019; Pabón-Caicedo et al., 2020) and amplified by thermodynamic increases in water vapour. Flood hazard from heavy rainfall is modulated by snowmelt (Section 8.2.3.1), vegetation characteristics (Murphy et al., 2020; Page et al., 2020) and direct human intervention (Section 8.2.3.4; FAQ 8.2) but also can be compounded by sea level rise (Sections 4.3.2.2; 9.6.4) in coastal and delta regions (Bevacqua et al., 2019; Ganguli and Merz, 2019; Eilander et al., 2020). Antecedent soil moisture conditions are an important modulator of flooding (Section 11.5.1) but become less important for smaller catchments and for more severe floods (Wasko and Nathan, 2019). Depleted soil moisture after more intense dry seasons (Section 8.2.2.1) can allow greater uptake of wet season rainfall before soils saturate. Since AR5, evidence confirms that more intense rainfall increases the proportion of runoff and reservoir recharge relative to infiltration into the soil (Eekhout et al., 2018; Yin et al., 2018). More intense but less frequent storms (Kendon et al., 2019) favour focused groundwater recharge through leakage from surface waters (Taylor et al., 2013b; Cuthbert et al., 2019a) and runoff and flash flooding where the percolation capacity of the soil is exceeded (Yin et al., 2018).

Increased severity of flooding on larger, more slowly-responding rivers is expected as precipitation accumulations increase during persistent wet events over a season. This can occur where atmospheric blocking patterns repeatedly steer extra tropical cyclones across large river catchments, as identified for Northern Hemisphere mid-latitudes and Asia (Takahashi et al., 2015; Lenggenhager et al., 2018; Pfleiderer et al., 2018; Zhou et al., 2018; Blöschl et al., 2019; Nikumbh et al., 2019; Zanardo et al., 2019), although groundwater flooding and antecedent conditions including soil moisture and snow melt also play a role (Muchan et al., 2015; Berghuijs et al., 2019). Increased atmospheric moisture amplifies the severity of these events when they occur in a warmer climate, yet drivers of change in the occurrence of blocking patterns, stationary waves and jet stream position are not well understood (Section 8.2.2.2, Cross Chapter Box 10.1).

In summary, there is *very high confidence* that heavy precipitation events will become more intense in a warming climate. There is *high confidence* that increased moisture and its convergence within extra-tropical and tropical cyclones and storms will increase rainfall totals during wet events at close to the 7% per ºC thermodynamic response, with *low confidence* of higher rates for sub-daily intensities. There is *medium confidence* that more intense but less frequent rainfall increases the proportion of rainfall leading to surface runoff and focused groundwater recharge from temporary water bodies. There is *low confidence* in how the frequency of flooding will change regionally as it is strongly dependent on catchment characteristics, antecedent conditions and how atmospheric circulation systems respond to climate change, which is less certain than thermodynamic drivers (Section 11.5). However, there is *high confidence* that increases in precipitation intensity and amount during very wet events (from sub-daily up to seasonal time-scales) will intensify severe flooding when these extremes occur.

8.2.3.3 Drivers of aridity and drought

Regional changes in aridity – broadly defined as a deficit of moisture – are expected to occur in response to anthropogenic forcings as a consequence of shifting precipitation patterns, warmer temperatures, changes in cloudiness (affecting solar radiation), declining snowpack, changes in winds and humidity, and vegetation cover (Figure 8.6). Evapotranspiration (see Annex VII: Glossary) is a key component of aridity, and is composed of two main processes: evaporation from soil, water and vegetation surfaces; and transpiration, the exchange of moisture between plants and atmosphere through plant stomata. On a global level, warmer temperatures increase evaporative demand in the atmosphere, and thus (assuming sufficient soil moisture is available) increase moisture loss from evapotranspiration (*high confidence*) (Dai et al., 2018; Vicente-Serrano et al., 2020b). On a regional level, aridity is further modulated by seasonal rainfall patterns, runoff, water storage, and interactions with vegetation.

Vegetation is a crucial interface between subsurface water storage (in soil moisture and groundwater) and the atmosphere. Plants alter evapotranspiration and the surface energy balance, and thus can have a large influence on regional aridity (Lemordant et al., 2018). SRCCL concluded there is *high confidence* that higher atmospheric CO_2 increases the ratio of plant CO_2 uptake to water loss (water-use efficiency; WUE) through the combined enhancement of photosynthesis and stomatal regulation (De Kauwe et al., 2013; Jones et al., 2013b; Deryng et al., 2016; Swann et al., 2016; Cheng et al., 2017; Knauer et al., 2017; Peters et al., 2018; Guerrieri et al., 2019) (see also Section 5.4.1). Modelling studies suggest that increasing WUE can partly counteract water losses from increased evaporative demand in a warmer atmosphere, potentially mitigating aridification (Milly and Dunne, 2016; Bonfils et al., 2017; Cook et al., 2018; Yang et al., 2018d). However, observational studies suggest that this effect may be counter-balanced by the increase in plant growth in response to elevated CO_2, which results in increased water consumption (De Kauwe et al., 2013; Donohue et al., 2013; Ukkola et al., 2016b; Yang et al., 2016; Guerrieri et al., 2019; Mankin et al., 2019; Singh et al., 2020a). In semi-arid regions, increased plant water consumption can reduce streamflow and exacerbate aridification (Ukkola et al., 2016b; Mankin et al., 2019; Singh et al., 2020a). Thus, there is *low confidence* that increased WUE in plants can counterbalance increased evaporative demand (Cross Chapter Box 5.1).

A drought is a period of abnormally dry weather that persists for long enough to cause a serious hydrological imbalance (Wilhite and Glantz, 1985; Wilhite, 2000; Cook et al., 2018) (see Annex VII: Glossary). Most droughts begin as persistent precipitation deficits (*meteorological drought*) that propagate over time into deficits in soil moisture, streamflow, and water storage (Figure 8.6), leading to a reduction in water supply (*hydrological drought*). Increased atmospheric evaporative demand increases plant water stress, leading to *agricultural and ecological drought* (Williams et al., 2013; Allen et al., 2015a; Anderegg et al., 2016; McDowell et al., 2016; Grossiord et al., 2020). Evaporative demand affects plants in two ways. It increases evapotranspiration, depleting soil moisture and stressing plants through lack of water (Teuling et al., 2013; Sperry et al., 2016), and also directly affects plant physiology, causing a decline in hydraulic conductance and carbon metabolism, leading to mortality (Breshears et al., 2013; Hartmann, 2015; McDowell and Allen, 2015; Fontes et al., 2018) (Figure 8.6). While droughts are traditionally viewed as "slow moving" disasters that typically take months or years to develop, rapidly evolving and often unpredictable *flash droughts* can also occur (Otkin et al., 2016, 2018). *Flash droughts* can develop within a few weeks, causing substantial disruption to agriculture and water resources (Pendergrass et al., 2020). Conversely, droughts that persist for a long time (usually a decade or more) are called *megadroughts*. Droughts span a large range of spatial and temporal scales, arise through a variety of climate system dynamics (e.g., internal atmospheric variability, ocean teleconnections), and can be amplified or alleviated by a variety of physical and biological processes. As such, droughts occupy a unique space within the framework of extreme climate and weather events, possessing no singular definition.

While the role of precipitation in droughts is obvious, other climatic drivers are also important, such as temperature, radiation, wind, and humidity (Figure 8.6). These factors have a strong influence on atmospheric evaporative demand, which affects evapotranspiration and soil moisture (Figure 8.6). In snow-dominated regions, high temperatures increase the fraction of precipitation falling as rain instead of snow and advance the timing of spring snowmelt (*high confidence*) (Vincent et al., 2015; Mote et al., 2016, 2018; Berg and Hall, 2017; Solander et al., 2018). This can result in lower than normal snowpack levels (*a snow drought*), and thus reduced streamflow, even if total precipitation is at or above normal for the cold season (Harpold et al., 2017). Plants also affect the severity of droughts by modulating evapotranspiration (Figure 8.6). As discussed above, the effect of elevated CO_2 on plants has the potential to both increase and reduce water loss through evapotranspiration via enhanced WUE and plant growth, respectively (Figure 8.6), but there is *low confidence* in whether one process dominates over another at the global scale.

Drought severity also depends on human activities and decision-making (AghaKouchak et al., 2015; Van Loon et al., 2016; Pendergrass et al., 2020). Societies have developed a variety of strategies to manipulate the water cycle to increase resiliency in the face of water scarcity, including irrigation, creation of artificial reservoirs, and groundwater pumping. While potentially buffering water resource capacity, in some cases these interventions may unexpectedly increase vulnerability (*medium confidence*). For example, while increased irrigation efficiency may ensure more water is available to crops, the corresponding reduction in runoff and subsurface recharge may exacerbate hydrologic drought (Grafton et al., 2018). Furthermore, while building dams and increasing surface reservoir capacity can boost water resources, they may actually increase drought vulnerability if demands rise to take advantage of the increased supply or if over-reliance on these surface reservoirs is encouraged (Di Baldassarre et al., 2018). Interactions between adaptation, vulnerability, and drought impacts are discussed further in WGII (Chapters 2 and 4).

In summary, there is *high confidence* that a warming climate drives an increase in atmospheric evaporative demand, decreasing available soil moisture. There is *high confidence* that higher atmospheric CO_2 increases plant water-use efficiency, but *low confidence* that this physiological effect can counterbalance water losses. Since drought can be defined in a number of ways, there are potentially different responses under a warming climate depending on drought type. Beyond a lack of precipitation, changes in evapotranspiration are critical components of drought, because these can lead to soil moisture declines (*high confidence*). Under very dry soil conditions, evapotranspiration becomes restricted and plants experience water stress in response to increased atmospheric demand (*medium confidence*). Human activities and decision-making have a critical impact on drought severity (*high confidence*).

[START FIGURE 8.6 HERE]

Figure 8.6: Climatic drivers of drought, effects on water availability, and impacts. Plus and minus signs denote the direction of change that drivers have on factors such as snowpack, evapotranspiration, soil moisture, and water storage. The three main types of drought are listed, along with some possible environmental and socioeconomic impacts of drought (bottom).

[END FIGURE 8.6 HERE]

8.2.3.4 Direct anthropogenic influence on the regional water cycle

Human activities influence the regional water cycle directly through modifying and exploiting stores and flows from rivers, lakes and ground water and by altering land cover characteristics. These actions alter surface energy and water balances through changes in permeability, surface albedo, evapotranspiration, surface roughness and leaf area. Direct redistribution of water by human activities for domestic, agricultural and industrial use of ~24,000 km^3 per year (Figure 8.1) is equivalent to half the global river discharge or double the global groundwater recharge each year (Abbott et al., 2019). Since the AR5, both modelling studies and observations have demonstrated that land use change can drive local and remote responses in precipitation and river flow by altering the surface energy balance, moisture advection and recycling, land-sea thermal contrast and associated wind patterns (Alter et al., 2015; Wey et al., 2015; De Vrese et al., 2016; Pei et al., 2016; Wang-Erlandsson et al., 2018; Vicente-Serrano et al., 2019). There is robust evidence that a warming climate combined with direct human demand for ground water will deplete ground water resources in already dry regions (Wada and Bierkens, 2014; D'Odorico et al., 2018; Jia et al., 2020).

SRCCL presented evidence that extraction of water from the ground or river systems and intensive irrigation increases evaporation and atmospheric water vapour locally (Jia et al., 2020; Mishra et al., 2020). Irrigation can explain declining groundwater storage in some regions, including north-western India and North America (Asoka et al., 2017; Ferguson et al., 2018b). Simulations spanning 1960-2010 indicate that ~30% of the present human water consumption is supplied from non-sustainable water resources (Wada and Bierkens, 2014). However, there is only limited evidence that groundwater extraction is lowering streamflow (Mukherjee et al., 2018; de Graaf et al., 2019). Model experiments show that irrigation can either aggravate or alleviate climate-induced changes of surface or sub-surface water (Leng et al., 2015). Widespread extraction of water from rivers can reduce flows and decrease the level and area of inland seas and lakes (Wurtsbaugh et al., 2017; Torres-Batlló et al., 2020; Wang et al., 2020d). Between 1985 and 2015, ~139,000 km^2 of inland water areas have become land, while creation of dams has converted about 95,000 km^2 of land to water, particularly in the Amazon and Tibetan Plateau (Donchyts et al., 2016). Direct management of river flow is comparable in magnitude to climate change effects for snow-fed rivers at a continental scale based on a global analysis and a study of 96 Canadian catchments (Tan and Gan, 2015; Arheimer et al., 2017).

SRCCL assessed with *medium confidence* that mean and extreme precipitation is increased over and downwind of urban areas (Jia et al., 2020). There is *medium confidence* that altered thermodynamic and aerodynamic properties of the land surface from urbanisation affects evaporation and increases precipitation over or downwind of cities (Box 10.3) due to altered stability and turbulence (Han et al., 2014; Pathirana et al., 2014; Jiang et al., 2016; D'Odorico et al., 2018; Sarangi et al., 2018; Boyaj et al., 2020), while reduced biogenic aerosol but increased anthropogenic aerosol emissions modify cloud microphysics and precipitation processes (Schmid and Niyogi, 2017; D'Odorico et al., 2018; Fan et al., 2020; Zheng et al., 2020)(Box 8.1). Urbanisation also decreases permeability of the surface, leading to increased surface runoff (Chen et al., 2017; Jia et al., 2020). Large-scale infrastructure, such as the construction and operation of dikes, weirs, and hydropower plants, also alters surface energy and moisture fluxes, potentially influencing the regional water cycle. *Limited* modelling *evidence* suggests that large-scale solar and wind farms can increase precipitation locally (over the Sahel and North America) when dynamic vegetation responses are represented (Li et al., 2018c; Pryor et al., 2020) with remote effects also possible (Lu et al., 2021).

Changes in land use from forest to agriculture can exert profound regional effects on the water cycle (FAQ 8.1) by modifying the surface energy balance and moisture recycling (Krishnan et al., 2016; Paul et al., 2016; Llopart et al., 2018; Singh et al., 2019). There is *medium evidence* from modelling and observations over the Amazon and East Africa that deforestation drives increased streamflow (Dos Santos et al., 2018; Guzha et al., 2018; Levy et al., 2018) but *limited evidence* that increases in global runoff due to deforestation are counterbalanced by decreases resulting from irrigation (Hoegh-Guldberg et al., 2019). Total Amazon deforestation drives large reductions in precipitation but with a 90% confidence range (-38 to +5%) based on 44 primarily pre-AR5 climate model simulations (Spracklen and Garcia-Carreras, 2015) with smaller reductions (-2.3 to -1.3%) attributed to observed Amazon deforestation up to 2010. Climate model development has reduced this uncertainty range but has not altered the median change (Lejeune et al., 2015). Large-scale global deforestation (20 million km^2) simulated by 9 CMIP6 models confirms a large range in precipitation amount reduction of -37±54 mm/yr over the deforested regions (Boysen et al., 2020). However, small-scale deforestation can increase precipitation locally (Lawrence and Vandecar, 2015). A 50–60% deforestation rate corresponded to a wet season delay of about one week and greater chance of dry spells of eight days or longer based on correlation analysis of rain gauge and land use data for South America (Leite-Filho et al., 2019). Forest and grassland fires can also modify hydrological response at the watershed scale (Havel et al., 2018). Afforestation or reforestation aimed at removing CO_2 from the atmosphere can also alter the water cycle at the regional scale (Section 8.4.3 and Cross-Chapter Box 5.1).

In summary, there is *high confidence* that land use change and water extraction for irrigation drive local, regional and remote responses in the water cycle. Large-scale deforestation is *likely* to decrease precipitation over the deforested regions but there is *low confidence* in the effects of limited deforestation. There is *medium confidence* that deforestation drives increased streamflow relative to the responses caused by climate change. There is *medium confidence* that urbanisation can increase local precipitation and runoff intensity. A warming climate combined with direct human demand for water is expected to deplete ground water resources in dry regions (*high confidence*).

[START BOX 8.1 HERE]

BOX 8.1: Role of anthropogenic aerosols in water cycle changes

Aerosols affect precipitation in two major pathways, by altering the shortwave and longwave radiation and influencing cloud microphysical properties.

Aerosol radiative effects on precipitation

Aerosols scatter and absorb solar radiation which reduces the energy available for surface evaporation and subsequent precipitation. In addition, cooling is incurred by the radiation that is reflected back to space directly by the aerosols and indirectly by the aerosol effect on cloud brightening. Northern Hemisphere (NH) station data indicate decreasing precipitation trends during 1950s-1980s, which have since partially recovered (Wild, 2012; Bonfils et al., 2020). These changes are attributable with *high confidence* to anthropogenic aerosol emissions from North America and Europe causing dimming through reduced surface solar radiation, which peaked during the late-1970s and partially recovered thereafter following improved air quality regulations (Section 6.2.1; Box 8.1, Figure 1).

[START BOX 8.1, FIGURE 1 HERE]

Box 8.1, Figure 1: Northern hemisphere surface downward radiation anomalies (Wm^{-2} ; a) and precipitation anomalies (mm/day ; b) for 1951–2014 for summer season (May–September) monsoon region (Polson et al. et al., 2014a) **from CMIP6 DAMIP experiments**. Observed solar radiation anomalies are from GEBA global data from 1961-2014 (Wild et al., 2017) and observed precipitation anomalies are from GPCC and CRU. CMIP6 multi-model mean anomalies are from all-forcings (ALL), greenhouse gas forcing (GHG) and anthropogenic aerosol forcing (AER) experiments. Anomalies are with respect to 1961–1990 and smoothed with a 11-year running mean. Red shading shows the ensemble spread of ALL forcing experiment (5%–95% range). Models are masked to the GPCC data set. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END BOX 8.1, FIGURE 1 HERE]

Dimming over the NH causes a relative cooling, compared to the SH, which induces a southward shift of the northern edge of the tropical rain belt (Allen et al., 2014; Brönnimann et al., 2015) (Section 3.3.2.2). CMIP5 simulations show that most of the cooling is caused by the aerosol cloud-mediated effect (Chung and Soden, 2017). Dimming also weakens monsoon flow and precipitation, offsetting or even overcoming the expected precipitation increase due to increased GHGs (Ayantika et al et al., 2021). The oceanic response to a weakened monsoon cross-equatorial flow can further weaken the South Asian monsoon through an amplifying feedback loop (Swapna et al., 2012; Krishnan et al., 2016; Patil et al., 2019) These processes partially explain (*medium confidence*) the southward shift of the NH tropical edge of the tropical rain belt from the 1950s to the 1980s (Allen et al., 2014; Brönnimann et al., 2015) and the severe drought in the Sahel that peaked in the mid-1980s (Rotstayn et al., 2002; Undorf et al., 2018). These processes also explain (*high confidence*) the observed decrease of southeast Asian Monsoon precipitation during the second half of the 20th century (Bollasina et al. et al., 2011; Sanap et al., 2015; Krishnan et al., 2016; Lau and Kim, 2017; Lin et al., 2018; Undorf et al., 2018) (Figure 8.7).

Absorption of solar radiation by anthropogenic aerosols such as black carbon warms the lower troposphere and increases moist static energy but also results in larger convection inhibition that suppresses light rainfall (Wang et al., 2013c) (Box 8.1, Figure 2). Release of aerosol-induced instability, often triggered by topographical barriers, produces intense rainfall, flooding (Fan et al., 2015; Lee et al., 2016) and severe convective storms (Saide et al., 2015) (*medium confidence*). In particular, aerosols induce intense convection at the Himalaya foothills during the pre-monsoon season, which generates a regional convergence there (*medium confidence)*. This mechanism is termed the “elevated heat pump hypothesis” (Lau and Kim, 2006; D’Errico et al., 2015).

[START BOX 8.1, FIGURE 2 HERE]

Box 8.1, Figure 2: Schematic depiction of the atmospheric effects of light absorbing aerosols on convection and cloud formation: (A) without and (B) with the presence of absorbing aerosols in the planetary boundary layer. The dashed and solid blue lines correspond to the vertical temperature profiles in the absence and presence of the absorbing aerosol layer, respectively, and the solid and dashed red lines denote the dry and moist adiabats, respectively. Absorbing aerosols result in an increasing temperature in the atmosphere but a reduced temperature at the surface. The reduced surface temperature and the increased temperature aloft lead to a larger negative energy associated with convective inhibition (-) and a higher convection condensation level (CCL) under the polluted conditions. On the other hand, the absorbing aerosol layer induces a larger convective available potential energy (+) above CCL, facilitating more intensive vertical development of clouds, if lifting is sufficient to overcome the larger convective inhibition. From (Wang et al., 2013).

[END BOX 8.1, FIGURE 2 HERE]

Aerosol cloud microphysical effects

Cloud droplets nucleate on pre-existing aerosol particles which act as cloud condensation nuclei (CCN). Anthropogenic aerosols add CCN, compared to a pristine background, and produce clouds with more numerous and smaller droplets, slower to coalesce into raindrops and to freeze into ice hydrometeors at temperatures below 0°C. Adding CCN suppresses light rainfall from shallow and short-lived clouds, but it is compensated by heavier rainfall from deep clouds. Adding aerosols to clouds in extremely clean air invigorates them by more efficient vapour condensation on the added drop surfaces (Koren et al., 2014; Fan et al., 2018). Clouds forming in more polluted air masses (hence with more numerous and smaller drops) need to grow deeper to initiate rain (Freud and Rosenfeld, 2012; Konwar et al., 2012; Campos Braga et al., 2017). This leads to larger amount of cloud water evaporating aloft while cooling and moistening the air there at the expense of the lower levels, which leads to convective invigoration (Dagan et al., 2017; Chua and Ming, 2020), followed by convergence, air mass destabilization and added rainfall in an amplifying feedback loop (Abbott and Cronin, 2021). In addition, delaying rain initiation until greater altitudes are reached transports more cloud water above the 0°C altitude and leads to additional release of latent heat of freezing and/or vapour deposition, which in combination with the added latent heat of condensation enhances the cloud updrafts (Fan et al., 2018). The stronger updrafts invigorate mixed phase precipitation and the resultant hail and cloud electrification (Rosenfeld et al., 2008a; Thornton et al., 2017). This includes the outer convective rainbands of tropical cyclones and there is *medium confidence* that air pollution enhances flood hazard associated with the outer rain bands at the expense of the inner rain bands (Wang et al., 2014; Zhao et al., 2018a; Souri et al., 2020).

The aerosol effect on invigoration and rainfall from deep convective clouds peaks at moderate levels (aerosol optical depth of 0.2 to 0.3), but reverses into suppression with more aerosols (Liu et al., 2019a). More generally, the microphysical aerosol-related processes often compensate or buffer each other (Stevens and Feingold, 2009). For example, suppressed rain by slowing drop coalescence enhances mixed phase precipitation. Therefore, despite the potentially large aerosol influence on the precipitation forming processes, the net outcome of aerosol microphysical effects on precipitation amount has generally *low confidence*, especially when evaluated with respect to the background of high natural variability in precipitation (Tao et al., 2012).

Ice nucleating particles (INP) aerosols initiate ice precipitation from persistent supercooled water clouds that have cloud droplets too small for efficient warm rain, or expedite mixed phase precipitation in short lived supercooled rain clouds (Creamean et al., 2013). Most INP are desert and soil dust particles, rather than air pollution aerosols (DeMott et al., 2010). Biogenic particles from terrestrial and marine origin are more rare, but important at temperatures above about -15°C (Murray et al., 2012; DeMott et al., 2016). Dust particles from long-range transport across the Pacific were found to enhance snow forming processes over the Sierra Nevada in California (Creamean et al., 2013; Fan et al., 2014). The impact of INP was demonstrated by glaciogenic cloud seeding experiments, which enhanced orographic supercooled clouds with *medium confidence* of success (French et al., 2018; Rauber et al., 2019; Friedrich et al., 2020). There are still major gaps in understanding the effects of INP mainly on deep convective clouds (Kanji et al., 2017; Stanford et al., 2017; Korolev et al., 2020)

[END BOX 8.1 HERE]

8.3 How is the water cycle changing and why?

This section focuses on the evaluation and attribution of past and recent water cycle changes using observational datasets, theoretical understanding and model simulations. Paleoclimate records and historical observations provide evidence for past water cycle changes caused both by natural variability and human activities (Haug et al., 2003; Buckley et al., 2010; Pederson et al., 2014). Key elements of the observed water cycle changes are assessed in this section, including flux and storage variations across the atmosphere, the continents and to a lesser extent the ocean and cryosphere, as well as related changes in large-scale atmospheric circulation and modes of variability. Particular emphasis is placed on assessing changes across regions and seasons (Box 8.2). Detailed regional assessments are presented in Chapters 10, 11, 12 and Atlas. Further information concerning large-scale observed water cycle changes and their attribution is available in Sections 2.3.1.3 and 3.3.2.

8.3.1 Observed water cycle changes based on multiple datasets

This section provides a process-based evaluation and a comprehensive assessment of observed water cycle changes by integrating multiple lines of evidence including paleoclimate data, historical datasets, theoretical understanding (Section 8.2) and model simulations.

8.3.1.1 Global water cycle intensity and P-E over land and oceans

The human influence on the global water cycle is often summarized as an intensification (Huntington, 2006; DeAngelis et al., 2015; Zhang et al., 2019c) or an overall strengthening which has been observed since at least 1980 (*high confidence*, see Chapter 2). There is however no unique definition of the global water cycle intensity (Trenberth, 2011; Ficklin et al., 2019; Sprenger et al., 2019). One simple metric is the global and annual mean amount of precipitation. Although an increase in global precipitation is consistent with physical expectations (Section 8.2.1), it has not yet been detected and attributed to human activities given large observational uncertainties and low signal-to-noise ratio (Section 8.3.3.2). Other metrics are more suitable to detect and attribute changes in the global water cycle, including the *likely* increase in global land precipitation since 1950 (Section 2.3.1.4) which is *likely* due to a human influence (Section 3.3.2.2).

The flux of fresh water between the ocean and atmosphere is determined by the difference between precipitation and evaporation (P-E). Evaporation is measured in very few locations across the global ocean, so that directly assessing P-E over the ocean is very challenging and relies on indirect reanalysis estimates (Robertson et al., 2020). AR5 presented *robust evidence* of an amplified oceanic pattern in P-E since the 1960s from both regional and global surface and subsurface salinity measurements and reanalyses. This pattern is consistent with our theoretical understanding of human induced changes in the water cycle, leading to the conclusion that these changes are *very likely* the result of anthropogenic forcings (Section 9.2.2.2).

In contrast, AR5 did not provide a conclusive assessment of observed changes in P-E over land. Continental P-E estimated from reanalyses and data-driven land-surface models indicate that interannual variations are linked to ENSO (Robertson et al., 2014, 2020). Increasing trends in P-E since 1979 based on land models are not statistically significant. Observations and models show evidence that P-E increases in the wet parts and decreases in the dry parts of tropical circulation systems, which shift in location seasonally and from year to year, with increases in seasonality since 1979 (Chou et al., 2013; Liu and Allan, 2013; Fu and Feng, 2014, see also Box 8.2).

In summary, a low signal-to-noise ratio, observational uncertainties and current data assimilation techniques limit the assessment of recent global trends in P-E over both land and ocean. It is *likely* that the global land P-E variations observed since the late 1970s were dominated by internal variability, mostly linked to ENSO teleconnections (*medium confidence*). In contrast, the attribution of changes in sea surface salinity (Section 3.5.2.2) suggests that it is *extremely likely* that human influence has contributed to the regional changes in P-E observed over the global ocean since the mid-20th century.

8.3.1.2 Water vapour and its transport

AR5 presented evidence of increases in global near-surface and tropospheric specific humidity since the 1970s but with *medium confidence* of a slowing of near-surface moistening trends over land associated with reduced relative humidity since the late 1990s. According to the AR5, radiosonde, Global Positioning System (GPS) and satellite observations of tropospheric water vapour indicate *very likely* increases at near global scales since the 1970s occurring at a rate that is generally consistent with the Clausius-Clapeyron relation (about 7% per ºC at low altitudes) and the observed atmospheric warming (Hartmann et al., 2013).

Since AR5, it is *very likely* that increases in global atmospheric water vapour are observed from in situ, satellite and reanalysis data (with *medium confidence* in the magnitude; Section 2.3.1.3). Satellite records show increases in upper tropospheric water vapour (constant relative humidity while temperatures have increased) since 1979 (Chung et al., 2014b; Blunden and Arndt, 2020), to which human influence has *likely* contributed (Section 3.3.2.1). Combined satellite and reanalysis estimates and CMIP6 atmosphere-only simulations (1988–2014) show global-mean precipitable water vapour increases of 6.7±0.3%/°C, very close to the Clausius Clapeyron rate (Allan et al., 2020). Satellite-based products show increases close to the Clausius-Clapeyron rate over the ice-free oceans (about 7 to 9 %/ºC; 1998-2008), but reanalysis estimates outside this range (Schröder et al., 2019) are an expected consequence of their changing observing systems (Allan et al., 2014; Parracho et al., 2018). Increases in precipitable water vapour are found over the central and sub-Arctic based on multiple reanalyses with some corroboration from sparse, in situ data (Vihma et al., 2016; Rinke et al., 2019; Nygård et al., 2020).

Declining near-surface relative humidity over land areas (e.g., United States, Mediterranean, south Asia, South America and southern Africa) is evident in surface observations (Willett et al., 2014, 2020; Dunn et al., 2017). This is consistent with a faster rate of warming over land than ocean (Byrne and O'Gorman, 2018) (see Sections 2.3.1.3 and 8.2.2.1). CMIP5 simulations underestimate the observed decreases in relative humidity over much of global land during 1979-2015 (Douville and Plazzotta, 2017; Dunn et al., 2017) even when observed SSTs are prescribed (-0.05 to -0.25 %/decade compared with an observed rate of -0.4 to -0.8 %/decade). It is not yet clear if this discrepancy is related to internal variability or can be explained by deficiencies in models (Vannière et al., 2019; Douville et al., 2020) or observations (Willett et al., 2014). Over the Northern Hemisphere mid-latitude continents, there is *medium confidence* that human influence has contributed to a decrease in near-surface relative humidity in summer (Sections 2.3.1.3 and 3.3.2.2).

Water vapour transport (or convergence) estimates from observations have substantial uncertainties even in regions of high quality radiosonde data. Consequently many studies use reanalyses for water transport estimates instead of instrumental observations. For example, increases in low-level (800-1000 hPa) moisture convergence into the tropical wet regime with a smaller outflow increase in the mid-troposphere (400-800 hPa) with warming was detected in one reanalysis (ERA-Interim) (Allan et al., 2014). Modelling evidence combined with statistical analysis demonstrate consistency between reanalysis moisture convergence and P-E over land (Robertson et al., 2016). Advances in reanalysis representation of atmospheric moisture and winds in addition to new observational isotope analysis have improved the ability to identify the main sources of water vapour for key continental regions and quantify the relative contributions from moisture advection and recycling (Gimeno et al., 2012; Van Der Ent et al., 2014; Joseph et al., 2016).

Observed changes in moisture transport can also arise from changes in atmospheric circulation as well as thermodynamics. For instance, moisture transport into the Arctic region estimated from reanalyses datasets is consistent with radiosonde data (Dufour et al., 2016)**,** with increases since 1979 linked to atmospheric circulation (Nygård et al., 2020). Moisture transport into the Eurasian Arctic was identified to increase by 2.6%/decade during 1948-2008 based on a reanalysis estimate (Zhang et al., 2013b). More intense moist intrusions associated with atmospheric rivers affecting the Arctic and Europe have been documented since 1979 but with a substantial influence from decadal internal variability (Ummenhofer et al., 2017; Mattingly et al., 2018). A recent strengthening of tropical circulation and associated moisture convergence has been idendified since around 2000 for the Amazon region (Arias et al., 2015; Barichivich et al., 2018; Espinoza et al., 2018a; Wang et al., 2018e). This was also strenghtened by increased moisture transport from the North Atlantic, driving more abundant latent heat release (Segura et al., 2020) and leading to an increased frequency of extreme floods in the northern Amazon (Barichivich et al., 2018; Heerspink et al., 2020). Overall, increased moisture transport has been linked to increased precipitation over wet tropical land areas (Gimeno et al., 2020) and to more extreme and persistent wet and dry weather events (Konapala et al., 2020) in many regions worldwide.

In summary, there is *high confidence* that human-caused global warming has led to an overall increase in water vapour and moisture transport throughout the troposphere, at least since the mid-1990s. In particular, there is *high confidence* that moisture transport into the Arctic has increased but only *medium confidence* in the attribution of such a trend to a human influence. There is *medium confidence* that human influence has contributed to a decrease in near-surface relative humidity over the Northern Hemisphere mid-latitude continents during summer (see also Sections 2.3.1.3 and 3.3.2.2).

8.3.1.3 Precipitation amount, frequency and intensity

This section assesses observed changes in precipitation at global and regional scales. Note that changes in precipitation seasonality are assessed in Box 8.2 and that changes in regional monsoons are assessed in section 8.3.2.4 where observed changes in both circulation and rainfall are considered. Further assessment of regional changes in precipitation is presented in Chapters 10, 12 and Atlas, while extreme precipitation is presented in Chapter 11.

AR5 concluded that it is *likely* there has been an overall increase in annual mean precipitation amount over mid-latitude land areas in the Northern Hemisphere, with *low confidence* since 1901, but *medium confidence* after 1951. There is further evidence of a faster increase since the 1980s (*medium confidence*) (Sections 2.3.1.3.4 and 3.3.2.1). Precipitation has increased from 1950 to 2018 over mid-high latitude Eurasia, most North America, southeastern South America, and northwestern Australia, while it has decreased over most of Africa, eastern Australia, the Mediterranean region, the Middle East, and parts of East Asia, central South America, and the Pacific coasts of Canada, as simulated by the CMIP5 multi-ensemble mean (Dai, 2021). Since AR5, there have been updates of several precipitation datasets, including satellite estimates, reanalysis and merged products (Adler et al., 2017; Roca, 2019). However, observational uncertainties remain an issue for assessing regional trends in seasonal or annual mean precipitation amount (Hegerl et al., 2015; Maidment et al., 2015; Sarojini et al., 2016), as well as the convective and stratiform types of precipitation (e.g., Ye et al., 2017). Precipitation trends at regional scales are dominated by internal variability across much of the world (Knutson and Zeng et al., 2018). Regional changes in precipitation amounts can also be obscured by contrasting responses to GHG versus aerosol forcings (Wu et al., 2013; Hegerl et al., 2015; Xie et al., 2016; Zhao and Suzuki, 2019; Zhao et al., 2020) and changes in precipitation intensity versus frequency (Shang et al., 2019).

Global and regional changes in precipitation frequency and intensity have been observed over recent decades. An analysis of 1875 rain gauge records worldwide over the period 1961–2018 indicates that there has been a general increase in the probability of precipitation exceeding 50 mm/day, mostly due to an overall boost in rain intensity (Benestad et al., 2019). Such changes in precipitation intensity and frequency have not been formally attributed to human activities, but are consistent with the heating effect of increasing CO_2 levels on the distribution of daily precipitation rates (Section 8.2.3.2) and with a distinct overall intensification of heavy precipitation events found in both observations and CMIP5 models, though with an underestimated magnitude (Fischer and Knutti, 2014). Beyond amplified precipitation extremes (Section 11.4.2), CMIP5 models also indicate that anthropogenic forcings have increased temporal variability of annual precipitation amount over land from 1950 to 2005, which is most pronounced in annual mean daily precipitation intensity (Konapala et al., 2017).

Anthropogenic aerosols can alter precipitation intensities both through radiative and microphysical effects (Box 8.1; Section 8.5.1.1.2). Precipitation suppression through aerosol microphysical effects has been observed in shallow cloud regimes over South America and the southeastern Atlantic, associated with local biomass burning (Andreae et al., 2004; Costantino and Bréon, 2010), and in industrial regions in Australia (Rosenfeld, 2000; Hewson et al., 2013; Heinzeller et al., 2016). In contrast, precipitation intensification through aerosol microphysical effects in deep convective clouds is seen in many regions such as the Amazon, southern United States, India, and Korea associated with anthropogenic aerosols from cities (Hewson et al., 2013; Fan et al., 2018; Lee et al., 2018b; Sarangi et al., 2018).

In the tropics, increases in precipitation amount are observed in convergence zones and decreases in the descending branches of the atmospheric circulation since 1979 (Chou et al., 2013; Liu and Allan, 2013; Gu et al., 2016; Polson et al., 2016; Polson and Hegerl, 2017), consistent with increased moisture transports with warming (Gimeno et al., 2020). Over tropical land areas, there is substantial variability in the "wet convergent regimes get wetter" and "dry divergent regimes get drier" pattern of trends observed since 1950 that are modulated by decadal changes in ENSO (Liu and Allan, 2013; Gu and Adler, 2018). CMIP6 models indicate an increased contrast between wet and dry regions in the tropics and subtropics (Schurer et al., 2020) (Figure 8.7). This provides further evidence that rainfall has increased in wet regimes, and slightly decreased in dry regimes over the period 1988-2019 (Figure 3.14). This greater contrast is primarily attributable to greenhouse gas forcings, although the observed trends are statistically larger than the model responses (Section 3.3.2.2).

Over the African continent, there are distinct precipitation trends observed in multiple datasets since the 1980s (Maidment et al., 2015; Nguyen et al., 2018b) (Figure 8.7). Increases in intense convective storms affecting the Sahel have been attributed to increased land-ocean temperature gradients (Taylor et al., 2017), enhanced by intense heating of the Sahara (Dong and Sutton, 2015) rather than thermodynamics (Section 8.2.2). Changes in Sahel rainfall, with reduced precipitation amounts from the 1960s to the 1980s and a subsequent recovery, are assessed in Section 8.3.2.4.3 and Section 10.4.2.1. In eastern Africa, decreasing precipitation amount (−2 to −7% per decade for 1983-2010) was reported for the March-to-May Long Rains season (Lyon and Dewitt, 2012; Viste et al., 2013; Liebmann et al., 2014; Maidment et al., 2015; Rowell et al., 2015) and evidence of a recovery since, with internal variability playing a large role in these decadal changes (Wainwright et al., 2019). In contrast, the second "Short Rains" season in Eastern Africa (October to December) does not exhibit significant precipitation trends (Rowell et al., 2015). Increases in annual southern Africa rainfall of 6-7% per decade during 1983-2010 are linked with the Pacific Decadal Oscillation (PDO) (Maidment et al., 2015).

Section 8.3.1.6 assesses changes in precipitation over the Mediterranean region and its connection with drought and aridity.

Rainfall increases have been observed over northern Australia since the 1950s, with most of the increases occurring in the north-west (Dey et al., 2018, 2019b; Dai, 2021) and decreases observed in the north-east (Li et al., 2012a) since the 1970s. In contrast, there has been a decline in rainfall over southern Australia related to changes in the intensification and position of the subtropical ridge (CSIRO and Bureau of Meteorology, 2015) and anthropogenic effects (Knutson and Zeng et al., 2018). The drying trend over southwest Australia is most pronounced during May–July, where rainfall has declined by 20% below the 1900–1969 average since 1970 and by about 28% since 2000 (Bureau of Meteorology and CSIRO, 2020).

Over South America, there is observational and paleoclimate evidence of declining precipitation amount during the past 50 years over the Altiplano and central Chile, primarily explained by the PDO but with at least 25% of the decline attributed to anthropogenic influence (Morales et al., 2012; Neukom et al., 2015; Boisier et al., 2016; Seager et al., 2019b; Garreaud et al., 2020). In contrast, a significant rainfall increase has been detected over the Peruvian-Bolivian Altiplano (from observational data and satellite-based estimations) since the 1980s (Imfeld et al., 2020; Segura et al., 2020) (Figure 8.7). Long-term (1902-2005) precipitation data indicate positive trends over southeastern South America and negative trends over the southern Andes, with at least a partial contribution from anthropogenic forcing (Gonzalez et al., 2014; Vera and Díaz, 2015; Díaz and Vera, 2017; Boisier et al., 2018; Knutson and Zeng et al., 2018) (see further assessment in Sections 10.4.2.2 and Atlas.7.2.2). The Peruvian Amazon has exhibited significant rainfall decreases during the dry season since 1980 (Lavado et al., 2013; Ronchail et al., 2018). Increases in wet season rainfall in the northern and central Amazon since the 1980s and decreases during the dry season in the southern Amazon (Barreiro et al., 2014; Gloor et al., 2015; Martín-Gómez and Barreiro, 2016; Espinoza et al., 2018a; Wang et al., 2018e; Haghtalab et al., 2020) are not explained by radiative forcing based on CMIP6 experiments (Figure 8.7) and trends are insignificant over longer periods since 1930 (Kumar et al., 2013) or more recently, since 1973 (Almeida et al., 2017) (see Section 8.3.2.4.5 for monsoon-related changes). For the tropical Andes region, trends in annual precipitation show heterogenous patterns, ranging between -4%/decade and +4%/decade in the north and south tropical Andes for a 30-year period at the end of the 20th century, although increases during 1965-1984 and decreases since 1984 have been registered in Bolivia (Carmona and Poveda, 2014; Pabón-Caicedo et al., 2020).

Over China, annual precipitation totals changed little from 1973 to 2016, but precipitation intensity significantly increased at a rate of 0.12 mm/day/decade, while the number of days with precipitation exceeding 0.1 mm/day significantly decreased at a rate of 0.9 days/decade (Shang et al., 2019). There is consistency in trend estimates during 1998-2015 over mainland China among satellite-based products and station data, which show increased precipitation amounts in autumn and winter and decreases in summer (Chen and Gao, 2018), consistent with a decreased intensity of East-Asian monsoon precipitation (Lin et al., 2014; Deng et al., 2018). Further assessment of precipitation changes over the South and Southeast Asian and the East Asian monsoon regions is presented in Section 8.3.2.4. An increasing trend in the frequency of heavy rainfall occurrences at the expense of low and moderate rainfall occurrences is found over central India (Krishnan et al., 2016; Roxy et al., 2017) and over eastern China with the latter due to increasing high aerosol levels (Qian et al., 2009; Guo et al., 2017a; Xu et al., 2017; Day et al., 2018), consistent with the effects of absorbing aerosol on stability and convective inhibition (Box 8.1).

Observed precipitation records since the early 1900s show increases in precipitation totals over central and northeastern North America that are attributable to anthropogenic warming but larger in magnitude than found in CMIP5 simulations (Knutson and Zeng et al., 2018; Guo et al., 2019). Decreases in precipitation amount over the central and southwestern United States and increases over the north-central United States during 1983-2015 (Cui et al., 2017; Nguyen et al., 2018b), are not clearly associated with forced responses in CMIP6 simulations (Figure 8.7; see also Section 10.4.2.3). Over Europe, precipitation trends since 1979 do not show coherence across datasets (Zolina et al., 2014; Nguyen et al., 2018b). Longer records since 1910 show increases for much of Scandinavia, northwestern Russia, and parts of northwestern Europe/UK and Iceland (Knutson and Zeng et al., 2018). Records since 1930 show increases of annual preciptation amount over western Russia (see also Section Atlas.8.2). Widespread increases in daily precipitation intensity appear clearly over regions with a high density of rain gauges, such as Europe and North America over the 1951-2014 period (Alexander, 2016). Observations during 1966-2016 over northern Eurasia show increases in the contribution of heavy convective showers to total precipitation by 1-2% on average (with local trends of up to 5%) for all seasons except for winter (Chernokulsky et al., 2019). Increases in convective precipitation intensity have been identified, particularly on sub-daily time-scales, using a range of modelling and observational data (Berg et al., 2013; Kanemaru et al., 2017; Pfahl et al., 2017).

Snowfall is an important component of precipitation in high-latitude and mountain watersheds. Reanalysis data indicate significant reductions in annual mean potential snowfall areas over the Northern Hemisphere land by 0.52 million km^2 per decade, with the largest decline over the Alps, with snow water equivalent reductions of about 20 mm per decade (Tamang et al., 2020). In the Tibetan Plateau, region-wide wintertime snowfall has increased but summer snowfall has decreased during the 1960-2014 period (Deng et al., 2017). State-of-the-art model simulations indicate reduced mean annual snowfall in the Arctic, despite the strong precipitation increase, mainly in summer and autum, when temperatures are close to the melting point (Bintanja and Andry, 2017).

[START FIGURE 8.7 HERE]

Figure 8.7: Linear trends in annual mean precipitation (mm/day per decade) for 1901-1984 (left) and 1985-2014 (right): (a & e) observational dataset, and the CMIP6 multi model ensemble mean historical simulations driven by, (b & f) all radiative forcings, (c & g) GHG only radiative forcings, (d & h) aerosol only radiative forcings experiment. Shade without grey cross correspond to the regions exceeding 10 % significant level. Grey crosses correspond to the regions not reaching the 10% statistical significant level. Nine CMIP6-DAMIP models have been used having at least 3 members. The ensemble mean is weighted per each model on the available and used members. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.7 HERE]

In summary, regional changes in precipitation amounts can be obscured by the contrasting responses to GHG and aerosol forcings across much of the 20th century and can be thus dominated by internal variability at decadal to multi-decadal timescales (*high confidence*). There is however a detectable increase in northern high-latitude annual precipitation over land which has been primarily driven by human-induced global warming (*high confidence*, see also Section 3.3.2). Human influence has strengthened the zonal mean precipitation contrast between the wet tropics and dry subtropics since the 1980s (*medium confidence*), although regional studies suggest a more complex precipitation response to evolving anthropogenic forcings. There is *high confidence* that daily mean precipitation intensities have increased since the mid-20th century in a majority of land regions with available observations and it is *likely* that such an increase is mainly due to GHG forcing (see Section 11.4). Section 8.3.2.4 assesses monsoon precipitation changes in detail.

8.3.1.4 Evapotranspiration

AR5 assessed that there was *medium confidence* that pan evaporation declined in most regions over the last 50 years, yet *medium confidence* that evapotranspiration increased from the early 1980s to the late 1990s. Since AR5, these conflicting observations have been attributed to internal variability and by the fact that evapotranspiration is less sensitive to trends in wind speed and is partly controlled by vegetation greening (Zhang et al., 2015a, 2016d; Zeng et al., 2018c). Observation-based estimates show a robust positive trend in global terrestrial evapotranspiration between the early 1980s and the early 2010s (Miralles et al., 2014b; Zeng et al., 2014, 2018c, Zhang et al., 2015a, 2016d). The rate of increase varies among datasets, with an ensemble mean terrestrial average rate of 7.6 ± 1.3 mm year^{-1} decade^{-1} for 1882–2011 (Zeng et al., 2018b). In addition, a decreasing trend in pan evaporation plateaued or reversed after the mid-1990s (Stephens et al., 2018a) has been reported as due to a shift from a dominant influence of wind speed to a dominant effect of water vapour pressure deficit, which has increased sharply since the 1990s (Yuan et al., 2019). The absence of a trend in evapotranspiration in the decade following 1998 was shown to be at least partly an episodic phenomenon associated with ENSO variability (Miralles et al., 2014b; Zhang et al., 2015a; Martens et al., 2018). Thus, there is *medium confidence* that the apparent pause in the increase in global evapotranspiration from 1998 to 2008 is mostly due to internal variability. In contrast to the AR5, there are now consistent trends in pan evaporation and evapotranspiration at the global scale, given the recent increase in both variables since the mid 1990s (*medium confidence*). Given the growing number of quantitative studies, there is *high confidence* that global terrestrial annual evapotranspiration has increased since the early 1980s.

Since AR5, the predominant contribution of transpiration to the observed trends in terrestrial evapotranspiration has been revisited and confirmed (Good et al., 2015; Wei et al., 2017). Using satellite and ecosystem models, Zhu et al., (2016) found a positive trend in leaf area index during 1982-2009, indicating that greening could contribute to the observed positive trend of evapotranspiration, in line with similar studies that focused on the 1981–2012 (Zhang et al., 2016d) and 1982–2013 (Zhang et al., 2015a) periods. Zeng et al. (2018) determined that the 8% global increase in satellite-observed leaf area index between the 1980s and the 2010s may explain an increase in evapotranspiration of 12.0+/-2.4mm/yr (about 55+/-25% of the total observed increase). Forzieri et al. (2020) estimated that the recent increase in leaf area index led to 3.66 ± 0.45 W/m^2 in latent heat flux (about 51 ± 6 mm/yr) and that the sensitivity of energy fluxes to leaf area index increased by about 20% over the 1982–2016 period. Overall, there is *medium confidence* that greening has contributed to the global increase in evapotranspiration since the 1980s.

Plant water use efficiency is expected to rise with CO_2 levels (*high confidence*, see Section 8.2.3.3 and Box 5.2), and can in theory counteract rising evapotranspiration in a warmer atmosphere (Section 8.2.3.3). However, observational studies suggest that this may not be the case in some ecosystems. For example, Frank et al. (2015) found that while the Water Use Efficency (WUE) increased in European forests across the 20th century, transpiration also increased due to more plant growth, a lengthened growing season, and increased evaporative demand. Likewise Guerrieri et al. (2019) observed that while WUE and photosynthesis increased in North American forests, stomatal conductance experienced only modest declines that were restricted to moisture-limited forests. Other studies further suggest that in many ecosystems increased WUE will not compensate for increased plant growth, amplifying declines in surface water availability (De Kauwe et al., 2013; Ukkola et al., 2016b; Singh et al., 2020a), while drought conditions can also offset the CO_2 fertilization effect and lead to a decline in WUE (Liu et al., 2020a). There is *low confidence* regarding the impact of plant physiological effects on observed trends in evapotranspiration.

An increasing number of studies have identified signals of attribution in the recent observed trends in evapotranspiration. Douville et al. (2013) found that the post-1960 rise in evapotranspiration in both the mid-latitudes and northern high latitudes was related to anthropogenic radiative forcing. An analysis of CMIP5 simulations suggests that anthropogenic forcing accounts for a large fraction of the global-mean evapotranspiration trend from 1982 to 2010 (Dong and Dai, 2017). Padrón et al., (2020) determined that increases in evapotranspiration were responsible for the majority of the anthropogenic pattern in dry-season water availability that dominates global trends since 1984. These findings are further supported by CMIP6 model results (Fig. 8.8) that show that the recent summertime increase in evapotranspiration in the northern mid-and-high latitudes is due to greenhouse gas forcing and decreasing anthropogenic aerosol emissions over Europe.

In summary, there is *high confidence* that terrestrial evapotranspiration has increased since the 1980s. There is *medium confidence* that this trend is driven by both increasing atmospheric water demand and vegetation greening, and *high confidence* that it can be partly attributed to anthropogenic forcing. There is *low confidence* about the extent to which increases in plant water use efficiency have influenced observed changes in evapotranspiration.

[START FIGURE 8.8 HERE]

Figure 8.8: Linear trends in annual mean evapotranspiration (mm/day per decade) for 1901-1984 (left) and 1985-2014 (right): (a & e) LMIP and observational dataset, and the CMIP6 multi model ensemble mean historical simulations driven by, (b & f) all radiative forcings, (c & g) GHG only radiative forcings, (d & h) aerosol only radiative forcings experiment. Shade without grey cross correspond to the regions exceeding 10 % significant level. Grey crosses correspond to the regions not reaching the 10% statistically significant level. Nine CMIP6-DAMIP models have been used having at least 3 members. The ensemble mean is weighted per each model on the available and used members. GLDAS is not available over the early 20th century so was replaced by a multi-model off-line reconstruction, LMIP, which is consistent with GLDAS over the recent period but may be less reliable over the early 20th century given larger uncertainties in the atmospheric forcings. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END OF FIGURE 8.8 HERE]

8.3.1.5 Runoff, streamflow and flooding

AR5 reported *low confidence* in the assessment of trends in global river discharge during the 20th century.

This is because many streamflow observations have been impacted by land use and dam construction, and the largest river basins worldwide differ in many characteristics, including geography and morphology. In regions with seasonal snow storage, AR5 WGII assessed that there is *robust evidence* and *high agreement* that warming has led to earlier spring discharge maxima and *robust evidence* of earlier breakup of Arctic river ice, as well as indications that warming has led to increased winter flows and decreased summer flows where streamflows are lower and that the observed increases in extreme precipitation led to greater probability of flooding at regional scales with *medium confidence*. SROCC found *robust evidence* and *high agreement* that discharge due to melting glaciers has already reached its maximum point and has begun declining with smaller glaciers, but only *low confidence* that anthropogenic climate change has already affected the frequency and magnitude of floods at the global scale.

Significant trends in streamflow and continental runoff were observed in 55 out of 200 large river basins during 1948-2012, with an even distribution of increasing and decreasing trends (Dai, 2016) (see also Section 2.3.1.3.6). A global detection and attribution study shows that the simulation of spatially heterogeneous historical trends in streamflow is consistent with observed trends only if anthropogenic forcings are considered (Gudmundsson et al., 2019). Section 3.3.2.3 assesses with *medium confidence* that anthropogenic climate change has altered regional and local streamflows, although a significant trend has not been observed in the global average (Sections 2.3.1.3.6 and 3.3.2.3). Multiple human-induced and natural drivers have been shown to play an important but variable role in observed regional trends of streamflow for several different areas (Fenta et al., 2017; Ficklin et al., 2018; Glas et al., 2019; Vicente-Serrano et al., 2019). For instance, decreasing runoff during the dry season have been observed over the Peruvian Amazon since the 1980s (Lavado et al., 2013; Ronchail et al., 2018). Up to 30–50% of the recent multi-decadal decline in streamflow across the Colorado River Basin can be attributed to anthropogenic warming and its impacts on snow and evapotranspiration (Woodhouse et al., 2016; McCabe et al., 2017; Udall and Overpeck, 2017; Xiao et al., 2018; Milly and Dunne, 2020). In the Upper Missouri River basin, Martin et al. (2020) found that warming temperatures have contributed to streamflow reductions since at least the late 20th century. Cold regions in the Northern Hemisphere have experienced an earlier occurrence of snowmelt floods, an overall increase in water availability and streamflow during winter, and a decrease in water availability and streamflow during the warm season (Aygün et al., 2019).

Some studies have suggested that dam construction and water withdrawals can be the dominant drivers in observed trends in streamflow amount (Wada et al., 2013). Regionally, land use and land cover changes have been identified as important factors for streamflow (Chen et al., 2020a). The impact of surface dimming from aerosol emissions on evaporation was identified as a discernible influence in Northern Hemisphere streamflows (Gedney et al., 2014). While changes in annual mean streamflow present a complicated picture, recent studies of changes in the timing of streamflow in snow-influenced basins continue to support a prominent influence from warming (Kang et al., 2016; Dudley et al., 2017; Kam et al., 2018). Global land runoff variations correlate significantly with ENSO variability (Miralles et al., 2014b; Schubert et al., 2016).

Observed changes in flooding are assessed in detail in Chapter 11 and are summarized as follows. For changes in the magnitude of peak flow, recent studies show strong spatial heterogeneity in the sign, size and significance of trends (Section 11.5.2.1). For changes in the frequency and magnitude of high flows, the conclusions remain limited by the large influence of water management (Section 11.5.2.2). For changes in timing of peak flows, recent studies further support observed changes in snowmelt-driven rivers (Section 11.5.2.3). Observed changes in runoff and flood magnitude cannot be explained by precipitation changes alone given the possible season- and region-dependent decreases in antecedent soil moisture and snowmelt which can partly offset the increase in precipitation intensity (Sharma et al., 2018), or the expected effect of urbanization and deforestation which can on the contrary amplify the runoff response (Chen et al., 2017; Abbott et al., 2019; Cavalcante et al., 2019). Simulations of mean and extreme river flows are consistent with the observations only when anthropogenic radiative forcing is considered (Gudmundsson and et al., 2021).

In summary, the assessment of observed trends in the magnitude of runoff, streamflow, and flooding remains challenging, due to the spatial heterogeneity of the signal and to multiple drivers. There is however *high confidence* that the amount and seasonality of peak flows have changed in snowmelt-driven rivers due to warming. There is also *high confidence* that land use change, water management and water withdrawals have altered the amount, seasonality, and variability of river discharge, especially in small and human-dominated catchments.

8.3.1.6 Aridity and drought

AR5 reported *low confidence* that changes in drought since the mid-20th century could be attributed to human influence, owing to observational uncertainties and difficulties in distinguishing decadal-scale variability from long-term trends. Changes in soil moisture, a metric of aridity, were not assessed thoroughly in AR5. Since AR5, new satellite products, land-surface reanalyses, and land-surface models have been used to document recent changes in soil moisture at the global scale. The science of detection and attribution has also progressed considerably (Trenberth et al., 2015; Easterling et al., 2016; Stott et al., 2016). Attribution efforts have further benefited from the increased use of paleoclimate information, which provides an important constraint on natural variability that is insufficiently sampled by short observational record (Cook et al., 2018; Kageyama et al., 2018).

Several studies have identified a persistent "fingerprint" of anthropogenic forcing in global trends in aridity spanning the last 120 years. Using a combination of tree ring data, CMIP5 model simulations, and reanalysis products, Marvel et al. (2019) determined that the dominant trend in aridity since 1900, characterized by drying in North and Central America and the Mediterranean, is detectable and attributable to external forcing from 1900–1949. This trend weakens from 1950–1975, possibly due to aerosol forcing (Marvel et al., 2019), but then emerges again from 1981 to present, although it is not detectable in the GLEAM nor MERRA-2 soil moisture reanalysis products. Likewise, Bonfils et al., (2020) investigated changes in precipitation, temperature and continental aridity in CMIP5 historical simulations and found that the dominant multivariate fingerprint, an amplification of wet-dry latitudinal patterns and progressive continental aridification, was associated with greenhouse gas emissions (Figure 8.9 a-d), and the second leading fingerprint was associated with anthropogenic aerosols (Figure 8.9 e-h). This study found that the anthropogenic greenhouse gas signal is statistically detectable in reanalyses over the 1950–2014 period (signal-to-noise ratio above 1.96). Gu et al. (2019) found that a global trend in declining soil moisture is detectable in the GLDAS-2 reanalysis product and is attributable to greenhouse gas forcing. Padrón et al. (2020) reconstructed the global patterns of dry season water availability from 1902–2014, and found it *extremely likely* (99% range) that trends in the last three decades of the analysis period could be attributed to anthropogenic forcing, mainly due to increases in evapotranspiration. It is *very likely* (>90% range) that anthropogenic forcing has affected global patterns of soil moisture over the 20th century.

On a regional scale, the robustness of trend attribution for drought and aridity varies widely. Key trends and their attributions are summarized here, while a complete regional assessment of observed trends in drought and aridity is in Chapter 11 (Sections 11.6.2, 12.3.2 and 12.4).

Several studies have analyzed CMIP5 and land surface models and detected a significant summertime drying trend in the Northern Hemisphere across the late 20th century that is attributable to anthropogenic forcings (Mueller and Zhang, 2016; Douville and Plazzotta, 2017). This trend is mainly driven by dryland areas such as the western United States and west-central Asia, where both reanalysis products and satellite data confirm there has been a persistent decline in soil moisture since 1990 (Liu et al., 2019d). In the western United States, snow deficits have *very likely* contributed to recent drying (Mote et al., 2018). Spring snow water equivalent across the Sierra Nevada Mountains reached a record low in 2015 (Margulis et al., 2016; Mote et al., 2016), possibly the lowest of the last five hundred years (Belmecheri et al., 2016). Over the longer California drought (2011–2015) anthropogenic warming alone reduced snowpack levels in the Sierras by 25% (Berg and Hall, 2017). The northwestern United States also experienced snow drought in 2015, despite near-normal levels of total cold season precipitation (Mote et al., 2016; Marlier et al., 2017). There is *high confidence* that anthropogenic warming contributed to these recent snow droughts (Belmecheri et al., 2016; Mote et al., 2016).

In the western United States, anthropogenic warming is amplifying drought and aridity by increasing evaporative demand and water loss to the atmosphere (Weiss et al., 2009; Overpeck, 2013; Cook et al., 2014;

Griffin and Anchukaitis, 2014; Williams et al., 2020). For the California drought between 2012–2014, Griffin and Anchukaitis (2014) used paleoclimate reconstructions to determine that while rainfall deficits were not unprecedented, record-high temperatures drove an exceptional decline in soil moisture relative to the last millennium. Williams et al. (2015) concluded that anthropogenic warming accounted for 8–27% of the these soil moisture deficits. Robeson (2015) estimated that the California drought was a 1-in-10,000 year event. Tree ring reconstructions indicate that prolonged megadroughts have occurred in the western United States throughout the last 1200 years (Cook et al., 2004, 2010, 2015a) forced by internal variability (Coats et al., 2016; Cook et al., 2016b). However, Williams et al. (2020) determined that 2000–2018 drought across the southwestern United States was the second driest 19-year period since 800 CE, and attributed nearly half the magnitude of this event to anthropogenic forcing (see also Section 10.4.2.3). Evidence for human signals in drought can also be found in western North American streamflow records, as noted above in Section 8.3.1.5. There is *high confidence* that anthropogenic forcing has contributed to recent droughts and drying trends in western North America.

Large areas of east-central Asia experienced drying in the early 2000s as a result of warmer temperatures, lower humidity, and declining soil moisture (Wei and Wang, 2013; Li et al., 2017d; Hessl et al., 2018). Paleoclimate data from the Mongolian plateau suggest that this recent central Asian drought exceeds the 900-year return interval, but is not unprecedented in the last 2060 years (Hessl et al., 2018). There is *low confidence* due to *limited evidence* that recent droughts in central Asia can be attributed to anthropogenic forcing.

The Mediterranean region has experienced notable changes in drought and aridity. A number of studies have identified a decline in precipitation since 1960 and attributed this to anthropogenic forcing (Hoerling et al., 2012; Gudmundsson and Seneviratne, 2016; Knutson and Zeng et al., 2018; Seager et al., 2019b). Kelley et al. (2015) showed that climate change caused a three-fold increase in the likelihood of the 2007–2010 meteorological drought in the eastern Mediterranean. However, historical trends in precipitation across the Mediterranean are spatially variable and contain substantial decadal variability, such that an anthropogenic influence may not be detectable in all areas (Zittis, 2018; Vicente-Serrano et al., 2020a). Records of soil moisture provide a clearer signal, indicating that higher temperatures and increased atmospheric demand have played a strong role in driving Mediterranean aridity (Vicente-Serrano et al., 2014). Hydrological modeling suggests that the recent decline in soil moisture in the Mediterranean is unprecedented in the last 250 years (Hanel et al., 2018). Paleoclimate evidence extends this view, additionally indicating that dryness in the Mediterranean is approaching an extreme condition compared to the last millennium (Markonis et al., 2018) and that the 15 year drought in the Levant (1998–2012) has an 89% likelihood of being the driest of the last 900 years (Cook et al., 2016a). Marvel et al. (2019) found that the Mediterranean region contributes strongly to the anthropogenic warming component of the global trend in aridity. There is *high confidence* that anthropogenic forcings are causing increased aridity and drought severity in the Mediterranean region.

Both central and northeastern Africa have experienced a decline in rainfall since about 1980 (Lyon and Dewitt, 2012; Lyon, 2014; Hua et al., 2016; Nicholson, 2017) (*high confidence*). In central Africa, the decline has been attributed to atmospheric responses to Indo-Pacific sea-surface temperature variability (Hua et al., 2018). In northeastern Africa, droughts have become longer and more intense in recent decades, continuing across rainy seasons (Hoell et al., 2017b; Nicholson, 2017), and this trend appears to be unusual in the context of the last 1500 years (Tierney et al., 2015). Knutson and Zeng (2018) attribute decreased annual precipitation over the Sudan to anthropogenic forcing, but other studies argue that the recent trend cannot yet be distinguished from natural variability, at least over parts of this region (Hoell et al., 2017b; Philip et al., 2018). There remains *low confidence* due to *limited evidence* that drying the northeastern Africa is attributable to human influence. In the West Cape region of South Africa, human influences increased the likelihood of the severe 2015–2017 drought by a factor of 3–6, depending on the analysis (Otto et al., 2018; Pascale et al., 2020). Anthropogenic forcing also contributed to the 2018 drought, mainly by increasing evapotranspiration (Nangombe et al., 2020). While some analysis of instrumental precipitation data in this region detect a slight long-term drying trend consistent with the simulated anthropogenic response (Seager et al., 2019b), there is strong multidecadal variability in the data (Wolski et al., 2020). However, an study of streamflow in southern Africa detected a significant decline (Gudmundsson et al., 2019) (see also Section 10.6.2). There is *medium confidence* in the long-term drying trend in this region and its attribution to anthropogenic forcing, and *medium confidence* that anthropogenic warming has contributed to recent severe drought events.

Several subtropical, semi-arid regions in the Southern Hemisphere have experienced long-term drying trends in the late 20th century. Southwestern South America (central Chile) experienced a multi-decadal decline in precipitation and streamflow culminating in a post-2010 megadrought that has been partly attributed to anthropogenic greenhouse gas emissions and ozone depletion (Boisier et al., 2016, 2018; Saurral et al., 2017; Knutson and Zeng et al., 2018; Seager et al., 2019b; Garreaud et al., 2020). There is *medium confidence* that drying in central Chile can be attributed to human influence. The tree-ring paleoclimate record demonstrates that the mid-century increase in exteme drought events in southern South America is unusual in the context of the last 600 years, suggesting an emerging influence of anthropogenic forcing (Morales et al., 2020).

There has been a 20% decrease in winter (May–July) rainfall in southwestern Australia since 1970, with the decline increasing to around 28% since 2000 (Delworth and Zeng, 2014; Bureau of Meteorology and CSIRO, 2020). There has also been a significant increase in the average intensity of seasonal droughts in the region since 1911in response to both lower precipitation and increased atmospheric evaporative demand (Gallant et al., 2013). Several studies attribute the precipitation declines in southwestern Australia to anthropogenic changes in GHG and ozone (Delworth and Zeng, 2014; Knutson and Zeng et al., 2018; Seager et al., 2019b). There is *high confidence* that the observed drying in southwestern Australia can be attributed to anthropogenic forcing.

In southeastern Australia, the average length of droughts have increased significantly, lasting between 10 and 69% longer than droughts during the first half of the 20th Century (Gallant et al., 2013). Paleoclimate reconstructions indicate a 97.1% probability that the decadal rainfall anomaly recorded during the 1997–2009 Millennium Drought in southeastern Australia was the worst experienced since 1783 (Gergis et al., 2012), and that the spatial extent and duration of cool season (April–September) rainfall anomalies were either very much below average or unprecedented over at least the last 400 years (Freund et al., 2017). Other paleoclimate studies suggest that the Millennium Drought in eastern Australia was not unusual in the context of natural variability reconstructed over the past millennium (Palmer et al., 2015; Cook et al., 2016c; Kiem et al., 2020). While there is currently *low confidence* that recent droughts in eastern Australia can be clearly attributed to human influences (Cai et al., 2014; Delworth and Zeng, 2014; Rauniyar and Power, 2020), there is emerging evidence that declines in April–October rainfall in southeastern Australia since the 1990s would not have been as large without the influence of increasing levels of atmospheric greenhouse gases (Rauniyar and Power, 2020).

In summary, it is *very likely* that anthropogenic factors have influenced global trends in aridity, mainly through competing changes in evapotranspiration and/or atmospheric evaporative demand due to anthropogenic emissions of GHG and aerosols. There is *high confidence* that the frequency and the severity of droughts has increased over the last decades in the Mediterranean, western North America, and southwestern Australia and that this can be attributed to anthropogenic warming. There is *medium confidence* that recent drying and severe droughts in southern Africa and southwestern South America can be attributed to human influence. In some regions of western North America and the Mediterranean, paleoclimate evidence suggests that recent warming has resulted in droughts that are of similar or greater intensity than those reconstructed over the last millennium (*medium confidence*).

[START FIGURE 8.9 HERE]

Figure 8.9: Spatial expressions (a-c; e-g) of the leading multivariate fingerprints of temperature (˚C), precipitation (mm/day), and aridity (CMI; the Climate Moisture Index) in CMIP5 historical simulations and the corresponding temporal evolution in both CMIP5 and reanalysis products (d, h). The first leading fingerprint is associated with greenhouse gas forcing (a-d) and the second leading fingerprint is associated with aerosol forcing (e-h). CMI is a dimensionless aridity indicator that combines precipitation and atmospheric evaporative demand. Figure after (Bonfils et al., 2020). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.9 HERE]

8.3.1.7 Freshwater reservoirs

8.3.1.7.1 Glaciers

AR5 and SROCC found, with *very high confidence,* a general decline in glaciers due to climate change in recent decades. There is *very high confidence* that during the decade 2010 to 2019 glaciers lost more mass than in any other decade since the beginning of the observational record (Chapter 2 section 2.3.2.3, Chapter 9 section 9.5.1. Human influence is *very likely* the main driver of the global, near-universal retreat of glaciers since the 1990s (Chapter 3, section 3.4.3.1). In Table 9.5, the contribution of glaciers to sea level rise for different periods is presented; in 1971-2018 glacier mass loss contributed 20.9 [10.0 to 31.7] mm or 22.1% of the sea level rise during that period. The highest mass loss rates are observed in the southern Andes, New Zealand, Alaska, Central Europe and Iceland while the largest mass loss are observed in Alaska, the periphery of Greenland and Arctic Canada (see section 9.5.1 and Figure 9.20). Predominantly, runoff from small glaciers such as in Canada has decreased because of glacier mass loss, while runoff from larger glaciers such as Alaska has typically increased (Bolch et al., 2010; Thomson et al., 2011; Tennant et al., 2012; WGMS, 2017; Huss and Hock, 2018). Asia contains the largest concentration of glaciers outside the polar regions where the total glacier mass change is −16.3 ± 3.5 Gt/yr over 2000-2016 with considerable intra-regional variability (Brun et al., 2017). Mass losses of glaciers in Asia between 2000 and 2018 are -19.0 +/-2.5 Gt/yr (Shean et al., 2020). The most negative changes were found in Nyainqentanglha with −4.0 ± 1.5 Gt/yr, while glaciers in Kunlun, northern Tibetan Plateau, slightly gained mass at 1.4 ± 0.8 Gt/yr.

There is some evidence that an increase of precipitation over high mountains can offset glacier ablation (melt) (Farinotti et al., 2020). However, this process has only been described from the Karakoram region in the northwestern Himalaya, where it is thought to be partly responsible to the advances of glacier changes in the last two decades, referred to as the 'Karakoram Anomaly' (Farinotti et al., 2020). In the Himalaya, Maurer et al. (2019) observed faster ice loss during 2000–2016 (–7.5 ± 2.3 Gt/yr) compared to 1975–2000 (–3.9 ± 2.2 Gt/yr). In the Southern Hemisphere, the rate of glacier mass lost in South America is estimated at 19.4 ± 0.6 Gt/yr based on surface elevation changes over 2000-2011, which include the North and South Patagonian Icefields of South America (Braun et al., 2019), and at −22.9 ± 5.9 Gt yr^{-1} over 2000-2018 (Dussaillant et al. 2019).

In summary, human-induced global warming has been the primary driver of a global glacier recession since the early 20th century (*high confidence*). Most glaciers have lost mass more rapidly since the 1960s and in an unprecedented way over the last decade, thereby contributing to increased glacier runoff, especially from larger glaciers until a maximum is reached, which tends to occur later in basins with larger glaciers and higher ice-cover fractions (*high confidence*).

8.3.1.7.2 Seasonal snow cover

AR5 assessed that Northern Hemisphere snow cover extent (SCE) has decreased since the late 1960s, especially in spring (*very high confidence*). This is confirmed by recent studies (Kunkel et al., 2016), (see also Chapter 2, Section 2.3.2.2). AR6 assesses that Northern Hemisphere spring snow cover has been decreasing since 1978 (*very high confidence*) and that this trend extends back to 1950 (*high confidence*) (see Section 9.5.3). Human-caused global warming is the dominant driver of this observed decline (Estilow et al., 2015) (see Section 3.4.2). Model simulations suggest that surface temperature responses at hemispheric/regional scales explain between 40% and 85% of the SCE trend variability (Mudryk, 2017). A decreasing trend in snowfall has also been detected in the Northern Hemisphere (Rupp et al., 2013) (Figure 8.1). Snowfall as a proportion of precipitation has decreased significantly in recent years (Berghuijs et al., 2014). However, a late 20th century increase in snowfall in West Antarctica observed in ice cores has been linked to a combination of factors including the anthropogenically forced deepening of the Amundsen Sea Low (Thomas et al., 2015, 2017).

Observations show a rapid recent decrease of spring SCE in Northern Hemisphere, mostly in Eurasia and

North America, closely linked to temperature change, e.g., March–April SCE is decreasing at 3.4%±1.1% $decade^{-1}$ (1979–2005) (Brown and Robinson, 2010; Hernández-Henríquez et al., 2015). An overall increasing annual trend of the Northern Hemisphere SCE since the late 1980s has been observed, in contrast to decreasing trends over 1960s-1980s that are dominated by the autumn and winter seasons (Barry and Gan, 2020). Such recent positive trends in snow cover extent are however at odds with other surface and satellite datasets and with the negative trends simulated by most CMIP5 and CMIP6 models (Mudryk et al., 2017, 2020). Hernández-Henríquez et al., (2015) also detected positive trends in October-November SCE in NOAA-CDR which are not replicated in other datasets (Section 9.5.3). Wu et al. (2018) found slower snowmelt rates over the Northern Hemisphere in 1980–2017, with higher ablation rates in locations with deep snow water equivalent (SWE), but due to the reduction of SWE in deep snowpacks, moderate/high ablation rates showed decreasing trends. Santolaria-Otín and Zolina, (2020) reported weak but significant decline in SCE in autumn over northern Eurasia and North America during 1979-2005, and similarly for spring, except for northern Siberia which showed higher spring SCE. Kapnick and Hall (2012) detected significant loss of spring mountain snowpack in western United States in 1950-2008. For Canada, extensive decreasing snow depths, SCE and duration were detected since mid-1970s, especially in western Canada during winter and spring (DeBeer et al., 2016). Berghuijs et al. (2014) show that across the continental United States, catchments with more snowfall than rainfall generally have higher mean streamflow, which will probably decrease with smaller fractions of precipitation falling as snow because of climate warming.

In summary, a decline in the springtime Northern Hemisphere snow cover extent, snow depth and duration has been observed since the late 1960s and has been attributed to human influence (*high confidence*). Depending on the region and season, there is *low-to-medium confidence* in the main drivers of snow cover changes, although various regions exhibit a shortening of the snow cover season which is consistent with global warming. A more detailed assessment of observed changes in seasonal snow cover is provided in Section 9.5.3.

8.3.1.7.3 Wetlands and lakes

Wetlands and lakes affect the climate through their impact on carbon and methane budgets (e.g. Saunois et al., 2016, Zhang et al., 2017) (Section 5.2.2) and on surface heat fluxes, with coupled weather and climate effects (e.g., Zhan et al., 2019). Although these features are also affected by human activities and by climate change, AR5 did not specifically report on wetlands and lakes.

Inventories of surface water bodies are not systematically produced at national or regional levels. However, assessments are undertaken at the global scale (Ramsar Convention on Wetlands, 2018). Merging observations from multiple satellite sensors makes it possible to detect surface water even under vegetation and clouds, over about 25 years but with low spatial resolution (Prigent et al., 2016). Most recent multi-satellite products from visible, infrared, and microwave measurements, estimate a surface water area of ~12-14 million km^2 (including permanent and transitory surfaces, e.g., Aires et al., 2018; Davidson et al., 2018), which is much higher than those provided by optical imagery (~3 million km^2). Inventories show a strong decrease in natural surface water of ~0.8% per year in total from 1970 to the present (Ramsar Convention on Wetlands, 2018) but the sites are not evenly distributed. Multi-satellite estimates show a strong inter-annual variability in surface water extent over the period 1992-2015 with no clear long-term trend (Prigent et al., 2020).

Human-made water bodies represent ~10% of the total continental water surfaces (Figure 8.1) (Ramsar Convention on Wetlands, 2018) and consist mainly of reservoirs and rice paddies. High resolution optical imagery over the period 1984-2015 (Donchyts et al., 2016; Pekel et al., 2016) shows that a net increase of ~ 0.1 million km^2 in artifical water surfaces, mainly due to the construction of reservoirs. Surfaces of rice paddies are also increasing, especially in South East Asia (Davidson et al., 2018).

In summary, there is *high confidence* that the extent of human-made surface water has increased over the 20^{th} and early 21^{st} century. In contrast, due to *limited agreement* in the observational records at the global scale, there is only *low confidence* in the observed decline of the natural surface water extent in recent years (see also SRCCL).

8.3.1.7.4 Groundwater

As the world's most widespread store of freshwater (Taylor et al., 2013b), groundwater is estimated to supply between a quarter and a third of the world's annual freshwater withdrawals to meet agricultural, industrial and domestic demands (Döll et al., 2012; Wada et al., 2014; Hanasaki et al., 2018). Attribution of changes in groundwater storage, observed locally through piezometry (Taylor et al., 2013b) or estimated from GRACE satellite measurements (Rodell et al., 2018) at regional scales (> 100.000 km^2) (Figure 8.10), is often complicated by non-climate influences that include land-use change (Favreau et al., 2009) and human withdrawals (Bierkens and Wada, 2019).

Following a global review of groundwater and climate change (Taylor et al., 2013b) and AR5 WGII, evidence of an association between heavy or extreme precipitation and groundwater recharge has continued to grow, especially in tropical (Asoka et al., 2018; Cuthbert et al., 2019a; Kotchoni et al., 2019) and sub-tropical regions (Meixner et al., 2016). Stable-isotope ratios of O and H at 14 of 15 sites across the tropics trace groundwater recharge to intensive monthly rainfall, commonly exceeding the ~70th intensity percentile (Jasechko and Taylor, 2015). Further, heavy rainfall recharging groundwater resources is often influenced by climate variability such as ENSO and PDO (Taylor et al., 2013c; Kuss and Gurdak, 2014; Asoka et al., 2017; Cuthbert et al., 2019b; Kolusu et al., 2019; Shamsudduha and Taylor, 2020). Additionally, increases in groundwater storage estimated from GRACE for 37 of the world's large-scale aquifer systems from 2002 to 2016 are generally found to result from episodic recharge associated with extreme (>90th percentile) annual precipitation.

The overall underestimation of precipitation intensities in global climate models (Wehner et al., 2010, 2020; Goswami and Goswami, 2017) and of their sensitivity to warming temperatures (Borodina et al., 2017) may lead to underestimates of their recharging effect on groundwater (Mileham et al., 2009; Cuthbert et al., 2019b). The limited ability of global climate models to represent key controls on regional rainfall variability like ENSO (Chen et al., 2020d) (Technical Annex VI, Section 3.7.3) may also underestimate observed recharge from such events that are of particular importance in drylands (Taylor et al., 2013c; Cuthbert et al., 2019b). Numerical representations of the impact of precipitation intensification on groundwater recharge in large-scale models remain constrained by the challenges of including key recharge pathways that consider preferential flowpaths in soils (Beven, 2018) and focused recharge through leakage from surface waters (Döll et al., 2014).

Increasing global freshwater withdrawals, primarily associated with the expansion of irrigated agriculture in drylands, have led to global groundwater depletion that has an estimated range of ~100 and ~300 km^3 yr^{-1} from hydrological models and volumetric-based calculations (Bierkens and Wada, 2019). The magnitude of this change is such that its estimated contribution to global sea-level rise is in the order of 0.3 to 0.9 mm yr^{-1} (Wada et al., 2010; Konikow, 2011; Döll et al., 2014; Pokhrel et al., 2015; de Graaf et al., 2017; Hanasaki et al., 2018)**.** Groundwater depletion has been observed regionally in the United States High Plains, California's Central Valley (Scanlon et al., 2012), northwest India (Rodell et al., 2009; Asoka et al., 2017), Upper Ganges in India (MacDonald et al., 2016), North China Plain (Feng et al., 2013), north-central Middle East region of Tigris-Euphrates-Western Iran (Voss et al., 2013), Central Asia (Hu et al., 2019), and North Africa (Bouchaou et al., 2013). The regional contribution of agricultural irrigation to groundwater depletion was previously highlighted by SRCCL but no formal assessment of observed changes in global or regional groundwater featured in AR5.

Quantification of changes in groundwater storage from GRACE is currently constrained by uncertainty in the estimation of changes in other terrestrial water stores using uncalibrated, global-scale Land Surface Models (Döll et al., 2014; Scanlon et al., 2018) and the limited duration of the period of GRACE observations (2002 to 2016). Centennial-scale piezometry in northwest India reveals that recent groundwater depletion traced by GRACE (Rodell et al., 2009; Chen et al., 2014), follows more than a century of groundwater accumulation through canal leakage (MacDonald et al., 2016). Further, groundwater depletion is often localised occurring below the footprint (200.000 km^2) of GRACE, as has been well demonstrated by detailed modelling studies in the California Central Valley (Scanlon et al., 2012) and North China Plain (Cao et al., 2016).

Climate variability and drought affect groundwater depletion mainly due to amplified groundwater withdrawals. For instance, the depletion rate in Central Valley aquifer in the USA from 2006 to 2010 is estimated to range from 6 to 8 $km^3 yr^{-1}$ using GRACE data (Scanlon et al., 2012). In India, Asoka et al. (2017) show contrasting trends in groundwater storage in the north (declining at 2 cm yr^{-1}) and south (increasing at 1-2 cm yr^{-1}) that is explained by variations in human withdrawals and precipitation linked to Indian Ocean sea surface temperature variability.

Changes in meltwater regimes from glaciers and seasonal snow packs tend to reduce the seasonal duration and magnitude of recharge (Tague and Grant, 2009). Aquifers in mountain valleys show shifts in the timing and magnitude of: (1) peak groundwater levels due to an earlier spring melt; and (2) low groundwater levels associated with lower baseflow periods (Allen et al., 2010; Dierauer et al., 2018; Hayashi, 2020). The effects of receding alpine glaciers on groundwater systems are not well understood but long-term loss of glacier storage is estimated to reduce summer baseflow (Gremaud et al., 2009). In permafrost regions, coupling between surface-water and groundwater systems may be particularly enhanced by warming (Lamontagne-Hallé et al., 2018; Lemieux et al., 2020). In areas of seasonal or perennial ground frost, increased recharge is expected despite a decrease in absolute snow volume (Okkonen & Kløve, 2011; Walvoord & Kurylyk, 2016).

Coastal aquifers are the interface between the oceanic and terrestrial hydrological systems. Global sea level rise (SLR) causes fresh–saline-water interfaces to move inland. The extent of seawater intrusion into coastal aquifers depends on a variety of factors including coastal topography, recharge, and groundwater abstraction from coastal aquifers (Comte et al., 2016). Modelling results suggest that the impact of SLR on seawater intrusion is negligible compared to that of groundwater abstraction (Ferguson & Gleeson, 2012; Yu & Michael, 2019). Coastal aquifers under very low hydraulic gradients, such as the Asian mega-deltas, are theoretically sensitive to SLR but, according to evidence from Akter et al. (2019) in the Ganges-Brahmaputra-Megna Basin, may be more severely and widely affected by changes in upstream river discharge. They argue further that saltwater inundation from storm surges will have the greatest localised effects.

In summary, there is *medium confidence* that increased precipitation intensities, partly due to human influence, have enhanced groundwater recharge, most notably in the tropics. There is *high confidence* that groundwater depletion has occurred since at least the start of the 21st century as a consequence of groundwater withdrawals for irrigation in some of the world's most productive agricultural areas in drylands (e.g. southern High Plains and California Central Valley in the USA, the North China Plain, northwest India).

[START FIGURE 8.10 HERE]

Figure 8.10: Trends in Terrestrial Water Storage (TWS) (in centimetres per year) obtained on the basis of GRACE observations from April 2002 to March 2016. The cause of the trend in each outlined study region is briefly explained and colour-coded by category. The trend map was smoothed with a 150-km-radius Gaussian filter for the purpose of visualization; however, all calculations were performed at the native 3° resolution of the data product. Figure from Rodell et al. (2018). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END OF FIGURE 8.10 HERE]

8.3.2 Observed variations in large-scale phenomena and regional variability

Observed changes in large-scale circulation indicators (Cross-Chapter Box 2.2) are assessed in Chapters 2 and 3 (see Sections 2.3.1.4 and 3.3.3). In this chaper we focus on the influence of regional scale teleconnection variabililty on the water cycle and the attribution of these circulation changes. While observed changes in modes of variability are assessed in Chapters 2 and 4 (see Sections 2.4 and 4.3.3), here focus on hydrological teleconnections of relevance to the water cycle.

8.3.2.1 Intertropical Convergence Zone (ITCZ) and tropical rain belts

AR5 concluded it is *likely* that the tropical belt, as delimited by the Hadley circulation, has widened since the 1970s. Observations in the satellite era indicate precipitation increases in the core of the Pacific ITCZ and decreases on the ITCZ margins (Gu et al., 2016; Su et al., 2017). As the satellite period has lengthened, observations have increasingly been used to assess trends in the ITCZ and tropical rain belt. Since AR5, significant narrowing and strengthening of the Pacific ITCZ after 1979 have been identified in atmospheric reanalyses (Wodzicki and Rapp, 2016), but no change in the ITCZ location (Byrne et al., 2018). Atmospheric model simulations suggest that with a narrower ITCZ, the subtropical jet becomes baroclinically unstable at a lower latitude and allows midlatitude eddies to propagate farther equatorward (Watt-Meyer and Frierson, 2019). Observational analyses also show that the ITCZ narrowing (Zhou et al., 2020) is associated with increased precipitation in the ITCZ core region that is strongly coupled to increasing Outgoing Longwave Radtion (OLR) in the expanding dry zones, particularly over land regions in the subtropics and midlatitudes (Lau and Tao, 2020). In addition, an eastward movement of the South Pacific Convergence Zone (SPCZ) between 1977 and 1999 has been reported, with associated significant precipitation trends in the South Pacific regions (Salinger et al., 2014).

ITCZ trends seen in satellites, precipitation measurements and reanalysis data are further supported by ocean surface-salinity observations. Long-term salinity observations show a freshening in the cores of the Atlantic and Pacific ITCZs and increased salinity on the ITCZ margins (Durack et al., 2010, 2012; Terray et al., 2012; Skliris et al., 2014). By investigating simultaneous changes in precipitation, temperature and continental aridity in CMIP5 historical simulations, Bonfils et al. (2020a) found a secondary signal (Figure 8.9, right column) characterized by a robust interhemispheric temperature contrast (Section 3.3.1.1), a latitudinal shift in the ITCZ (in accordance with the theory of cross-equatorial energy transport; Section 8.2.2.2), and changes in aridity in the Sahel (section 8.3.1.6). These forced changes are statistically detectable in reanalyses datasets over the 1950-2014 period at the 95% confidence level.

Reconstructions in the Sahel (Carré et al., 2019) and Belize (Ridley et al., 2015) support the southward displacement of the tropical rain belt since 1850 and the narrowing trend of the tropical rainbelt detected in observations (Rotstayn et al., 2002; Hwang et al., 2013). Decreasing precipitation trends in the Northern Hemisphere during the 1950s-1980s have been attributed to anthropogenic aerosol emissions from North America and Europe, which peaked during the late1970s and declined thereafter following improved air quality regulations, causing dimming (brightening) through reduced (increased) surface solar radiation (Box 8.1 Figure 1), in agreement with model simulations (Chiang et al., 2013; Hwang et al., 2013). This is consistent with energetic constraints where tropical precipitation shifts are anti-correlated with cross-equatorial energy transport (Section 6.3.3, Box 8.1). It also provides a physical mechanism for the severe drought in the Sahel that peaked in the mid-1980s (Sections 8.3.2.4.3 and 10.4.2.1) and the southward shift of the Northern Hemisphere tropical edge from the 1950s to the 1980s (Allen et al., 2014; Brönnimann et al., 2015). However, CMIP5 and CMIP6 models still exhibit strong biases in the representation the ITCZ, such as the simulation of a double ITCZ (Oueslati and Bellon, 2015; Adam et al., 2018; Tian and Dong, 2020). The impacts of aerosols and volcanic activity on the position of the ITCZ have been investigated but changes are difficult to characterize from observations (Friedman et al., 2013; Haywood et al., 2013b; Iles et al., 2014; Colose et al., 2016; Chung and Soden, 2017) (see Section 6.3.3.2). Such systematic shifts of the ITCZ can have important regional impacts like changes in precipitation (Figure 8.9).

In summary, there is *medium confidence* that the tropical rain belts over the oceans have been narrowing and strengthening in recent decades, leading to increased precipitation in the ITCZ core region (see also Section 8.2.2.2). Decreasing precipitation trends in the Northern Hemisphere during the 1950s-1980s have been attributed to anthropogenic aerosol emissions from North America and Europe (*high confidence)*.

8.3.2.2 Hadley circulation and subtropical belt

AR5 reported *low confidence* in trends in the strength of the Hadley Circulation (HC) due to uncertainties in reanalyses but *high confidence* on the widening of the tropical belt since 1979. In AR6, Chapter 2 (see section 2.3.1.4.1) states that the HC has *very likely* widened and strengthened since at least the 1980s, mostly in the Northern Hemisphere (*medium confidence*).

The poleward shift of the HC is closely related to migration of the location of tropical cyclone trajectories in both hemispheres (Studholme and Gulev, 2018; Sharmila and Walsh, 2018), with a *very likely* poleward shift over the western North Pacific Oceans since the 1940s (Section 11.7.1.2). Moreover, the Western North Pacific Subtropical High has extended westward since the 1970s, resulting in a monsoon rain band shift over China, with excessive rainfall along the middle and lower reaches of the Yangtze River valley along ~ 30°N over eastern China. At the same time, the effect of anthropogenic aerosols dominated the response to GHG increases over East Asia, resulting in a weakening of the East Asian summer monsoon and causing a drying trend in northeastern China (Hu, 2003; Yu and Zhou, 2007; Wang et al., 2013b; Li et al., 2016c; Lau and Kim, 2017) and northern parts of South Asia (Preethi et al., 2017) (see also Section 8.3.2.4.2). During 1977-2007, the precipitation variability over the eastern United States increased due to changes in the intensity and position of the western ridge of the North Atlantic Subtropical High (Li et al., 2011; Diem, 2013).

In the Southern Hemisphere, the HC expansion has been associated with both the intensification and poleward shift of the subtropical high pressure belt (Nguyen et al., 2015), with consequences for precipitation amount over Africa, Australia, South America, and subtropical Pacific islands (Cai et al., 2012; Grose et al., 2015; Nguyen et al., 2015; Sharmila and Walsh, 2018; McGree et al., 2019). The subtropical ridge in Australia has intensified significantly since 1970, with marked declines observed in April to October rainfall across southeastern and southwestern Australia (Timbal and Drosdowsky, 2013).

The local tropical edges of the meridional overturning cells (as diagnosed from the horizontally divergent wind) are more closely associated with hydroclimate variations than the subtropical ridge (Staten et al., 2019). Poleward expansion of the tropical belt strongly contributes to precipitation decline in the poleward edge of the subtropics (Cai et al., 2012; Scheff and Frierson, 2012; Timbal and Drosdowsky, 2013; He and Soden, 2017; Nguyen et al., 2018a; Tang et al., 2018), although recent modelling evidence suggests that subtropical precipitation declines are a response to direct CO_2 radiative forcing mainly over ocean, irrespective of the HC expansion (He and Soden, 2017). Both reanalyses datasets and climate model simulations suggest that the HC expansion is not associated with widespread, zonally symmetric subtropical drying over land (Schmidt and Grise, 2017).

Since AR5, an improved understanding of the key drivers of the recent HC expansion has been achieved, identifying the role of both internal variability and anthropogenic climate change. Part of the recent expansion (1979-2005) of the HC has been driven by a swing from warm to cold phase of the Pacific Decadal Variability (PDV) (Meehl et al., 2016; Grise et al., 2019). The presence of large multidecadal variability in 20th century reanalyses means there is *limited evidence* on the human influence on the recent HC strengthening, yet the southward shift of the southern edge and widening of the Southern Hemisphere HC appeared as robust features in all reanalysis datasets, and their trends have accelerated during 1979-2010 (D'Agostino and Lionello, 2017). As assessed in Section 3.3.3.1, GHG increases and stratospheric ozone depletion have contributed to the expansion of the zonal mean HC in the Southern Hemisphere since around 1980, and the expansion of the Northern Hemisphere HC has not exceeded the range of internal variability (*medium confidence*). Moreover, Antarctic ozone depletion can cause a poleward shift in the Southern Hemisphere midlatitude jet and HC (Sections 3.3.3 and 6.3.3.2). Further assessment of the attribution of recently observed changes in the HC extent and intensity is found in Section 3.3.3.1.

In summary, it is *very likely* that the recent Hadley Circulation (HC) expansion was associated with poleward shifts of tropical cyclone tracks over the western North Pacific Ocean since the 1940s, and of extratropical storm tracks in the Southern Hemisphere since the 1970s. Changes to the HC in the Northern Hemisphere may have contributed to subtropical drying and a poleward expansion of aridity during the boreal summer, but there is *low confidence* due to *limited evidence*. GHG increases and stratospheric ozone depletion have contributed to expansion of the zonal mean HC in the Southern Hemisphere since around 1970, while the expansion of the Northern Hemisphere HC has not exceeded the range of internal variability (*medium confidence*).

8.3.2.3 Walker circulation

AR5 concluded that the long-term weakening of the Pacific Walker Circulation (WC) from the late 19th century to the 1990s has been largely offset by a recent strengthening (*high confidence*), though with *low confidence* in trends of the WC strength due to reanalysis uncertainties and large natural variability. The observed trends in the WC since 1980 are consistent with a *very likely* WC strengthening in the Pacific, similar to a La Niña pattern, with *medium confidence* in the magnitude of these changes due to differences between satellite observations and reanalyses.

The causes of the observed strengthening of the Pacific WC during 1980-2014 are not well understood due to competing influences from individual external forcings and since this strengthening is outside the range of variability simulated in coupled models (*medium confidence*), as assessed in Chapter 3 (see Section 3.3.3.1). Recent strengthening in the WC has been linked with internal variability (Chung et al., 2019), although one study argues that it could be a response forced by GHG that models do not capture because of common SST biases in the equatorial Pacific (Seager et al., 2019a). It could be also related to an interbasin thermostat mechanism whereby the human-induced Indian Ocean warming emerged earlier than in the tropical Pacific (Zhang et al., 2018b) and induced a transient strengthening of the zonal sea level pressure gradient and easterly trades in the tropical Pacific (Zhang et al., 2019a).

The weakening of the WC observed during most of the 20th century is associated with reductions in land rainfall over the Maritime Continent during 1950-1999 (Tokinaga et al., 2012; Yoden et al., 2017). In contrast, the recent strengthening of the WC has been associated with an intensification of extreme flooding (Barichivich et al., 2018) and an increased frequency of wet days (Espinoza et al., 2016, 2018a) over the northwestern Amazon, increased precipitation in South America (Yim et al., 2017), reduced precipitation over eastern Africa (Williams and Funk, 2011; Lyon and Dewitt, 2012), and increased rainfall in southern Africa (Maidment et al., 2015). Internal variability has been shown to have a dominant role in the recent strengthening of the WC (Chung et al., 2019).

In summary, there is *high confidence* that changes in the Pacific Walker Circulation (WC) are associated with changes in the water cycle over regions like the Maritime Continent, South America and Africa. It is *very likely* that the WC has strengthened in the Pacific since the 1980s, with *medium confidence* that this strengthening is within the range of internal variability.

8.3.2.4 Monsoons

AR5 reported *low confidence* in the attribution of changes in monsoons to human influences, although a detailed attribution assessment of the observed changes in the regional monsoons was not presented.

Large human populations in the monsoon regions of the world heavily depend on freshwater supply for agriculture, water resources, industry, transport and various socio-economic activities. The effects of GHG forcing combined with water vapour feedback (Allen et al., 2015b; Dong and Sutton, 2015; Evan et al., 2015; Dunning et al., 2018) and cloud feedbacks (Stephens et al., 2015; Potter et al., 2017) are fundamental to monsoon precipitation changes in a warming world. Since AR5 there has been improved understanding of precipitation changes associated with regional monsoons. Sections 2.3.1.4.2 and 3.3.3.2 provide an assessment of observed changes and attribution for the global monsoon. Here we provide an assessment of the observed changes in regional monsoons (see Annex V and Figure 8.11) and underlying causes. In AR6, the definition of regional monsoons slightly differs from AR5 and the rationale for it is provided in Annex V (see also Annex VII: Glossary). Specific examples of regional monsoons are discussed further in section 10.4.2, from the perspective of climate change attribution and in section 10.6.3, from the viewpoint of constructing regional climate messages.

[START FIGURE 8.11 HERE]

Figure 8.11: Regional monsoon precipitation changes from observations and model attribution. Precipitation changes during 1951-2014 are shown as least-square linear trends in box-whisker plots (first and fourth rows) over the six regional monsoons, i.e., North American monsoon (NAmerM, Jul-Aug-Sep), West African monsoon (WAfriM, Jun-Jul-Aug-Sep), South and Southeast Asian monsoon (SAsiaM, Jun-Jul-Aug-Sep), East Asian monsoon (EAsiaM, Jun-Jul-Aug), South American monsoon (SAmerM, Dec-Jan-Feb), Australian and Maritime Continent monsoon (AusMCM, Dec-Jan-Feb), and over the two land domains (i.e. equatorial America (EqAmer, Jun-Jul-Aug) and South Africa (SAfri, Dec-Jan-Feb), as identified in the map shown in the middle and as described in Annex V. Precipitation changes are computed from observations and from DAMIP CMIP6 experiments over the historical period with all-forcing (ALL), GHG-only forcing (GHG), Aerosol-only (AER) and Natural (NAT) forcings prescribed. Observations are based on the CRU (light green) and GPCC (light blue) datasets and the APHRODITE (light orange) dataset for SAsiaM and EAsiaM. CMIP6 simulations are taken from nine CMIP6 models contributing to DAMIP, with at least 3 members. Ensembles are weight-averaged for the respective model ensemble size. Observed trends are shown as coloured cirles and the simulated trends from the CMIP6 multi-model experiments are shown as box-whisker plots. Precipitation anomaly time-series are shown in the second and third row. The thick black line is the multi-model ensemble-mean precipitation anomaly time-series from the ALL experiment and the grey shading shows the spread across the multi-model ensembles. A 11-year running mean has been applied on the precipitation anomaly time-series prior to calculating the multi-model ensemble mean. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.11 HERE]

8.3.2.4.1 South and Southeast Asian Monsoon

AR5 reported a decreasing trend of global land monsoon precipitation over the last half-century, with primary contributions from the weakened summer monsoon systems in the Northern Hemisphere. Since AR5, several studies have documented long-term variations and changes in the South and Southeast Asian Summer Monsoon (SAsiaM) rainfall. The SAsiaM strengthened during past periods of enhanced summer insolation in the Northern Hemisphere, such as the early-to-mid Holocene warm period around 9000 to 6000 years BP (Masson-Delmotte et al., 2013; Mohtadi et al., 2016; Braconnot et al., 2019) and weakened during cold periods (*high confidence*), such as the Last Glacial Maximum (LGM) and Younger Dryas (Shakun et al., 2007; Cheng et al., 2012; Dutt et al., 2015; Chandana et al., 2018; Hong et al., 2018; Zhang et al., 2018a). These long time-scale changes in monsoon intensity are tightly linked to orbital forcing and changes in high-latitude climate (Braconnot et al., 2008; Battisti et al., 2014; Araya-Melo et al., 2015; Rachmayani et al., 2016; Bosmans et al., 2018; Zhang et al., 2018a). A weakening trend of the SAsiaM during the last 200 years has been documented based on tree ring oxygen isotope chronology from the northern Indian subcontinent (Xu et al., 2018) and Southeast Asia (Xu et al., 2013), oxygen isotopes in speleothems from northern India (Sinha et al., 2015), and tree ring width chronologies from the Indian core monsoon region (Shi et al., 2017). Nevertheless, the detection of century-long decreases in regional monsoon rainfall is obscured by the presence of multidecadal time-scale precipitation variations (Turner and Annamalai, 2012; Knutson and Zeng et al., 2018) which are evident in long-term rain guage records extending back to the early 1800s (Sontakke et al., 2008) and emerge in long-term climate simulations (Braconnot et al., 2019).

A significant decline in summer monsoon precipitation is observed over India since the mid-20th century, which is accompanied by a weakening of the large-scale monsoon circulation (Mishra et al. et al., 2012; Abish et al., 2013; Krishnan et al., 2013, 2016; Saha et al., 2014; Roxy et al., 2015; Guhathakurta et al. et al., 2017; Samanta et al. et al., 2020). This precipitation decline is corroborated by a decreasing trend in the frequency of monsoon depressions that form over Bay of Bengal (Prajeesh et al., 2013; Vishnu et al., 2016), an increasing trend in the frequency and duration of monsoon breaks or 'dry spells' (Singh et al., 2014), significant decreases in soil moisture and increases in drought severity across different parts of India post-

1950 (Niranjan Kumar et al., 2013; Ramarao et al., 2015; Krishnan et al., 2016; Ramarao et al., 2018; Ganeshi et al., 2020; Mujumdar et al., 2020). While recent studies have reported an apparent recovery of the Indian summer monsoon over a relatively short period since 2003 (Jin and Wang, 2017; Hari et al., 2020), long-term trends for the period 1951-2015 indicate an overall decrease in the regional monsoon precipitation (Kulkarni et al., 2020; Ayantika et al et al., 2021). A case study on the Indian summer monsoon is provided in Section 10.6.3.

Evidence from several climate modelling studies indicates that the observed decrease in the regional monsoon precipitation during the second half of the 20th century is dominated by the radiative effects of Northern Hemisphere anthropogenic aerosols, with smaller contributions due to volcanic aerosols from the Mount Pinatubo (1991) and El Chichon (1982) eruptions (Bollasina et al. et al., 2011; Polson et al. et al., 2014a; Sanap et al., 2015; Krishnan et al., 2016; Liu et al., 2016; Lau and Kim, 2017; Undorf et al., 2018; Undorf et al. et al., 2018; Lin et al., 2018; Takahashi et al., 2018; Patil et al., 2019; Singh et al., 2020b) (Box 8.1 Figure 1, Figure 8.11). Land-use changes over South and Southeast Asia and the rapid warming trend of the equatorial Indian Ocean during the recent few decades also appear to have contributed to the observed decrease in monsoon precipitation (Roxy et al., 2015; Krishnan et al., 2016; Singh, 2016). Overall, the magnitude of the precipitation response to anthropogenic forcing exhibits large spread across CMIP5 models pointing to the strong internal variability of the regional monsoon (Saha et al., 2014; Salzmann et al., 2014; Sinha et al., 2015), including variations linked to phase changes of the Pacific interdecadal variability (Section AVI.2.6) (Huang et al., 2020b), uncertainties in representing aerosol-cloud interactions (Takahashi et al., 2018), and the effects of local versus remote aerosol forcing (Bollasina et al. et al., 2014; Polson et al. et al., 2014b; Undorf et al., 2018). CMIP3 and CMIP5 models do not accurately reproduce the observed seasonal cycle of precipitation over the major river basins of South and Southeast Asia, limiting the attribution of observed regional hydroclimatic changes (Hasson, 2014; Hasson et al., 2016; Biasutti, 2019). While warm rain processes and organized convection are known to dominate the heavy orographic monsoon rainfall over the Western Ghats mountains (Shige et al., 2017; Choudhury et al., 2018), in various parts of India (Konwar et al., 2012) and East Asia (Section 11.7.3.1), there are uncertainties in representing the regional physical processes of the monsoon environment, including cloud-aerosol interactions (Sarangi et al., 2017), land-atmosphere (e.g., Barton et al., 2020) and ocean-atmosphere coupling (Annamalai et al., 2017), in state-of-the-art climate models (See also Section 8.5.1).

In summary, there is *high confidence* in observational evidence for a weakening of the SAsiaM in the second half of the 20th century. Results from climate models indicate that anthropogenic aerosol forcing has dominated the recent decrease in summer monsoon precipitation, as opposed to the expected intensification due to GHG forcing (*high confidence*). On paleoclimate timescales, the SAsiaM strengthened in response to enhanced summer warming in the NH during the early-to-mid Holocene, while it weakened during cold intervals (*high confidence*). These changes are tightly linked to orbital forcing and changes in high-latitude climate (*medium confidence*).

8.3.2.4.2 East Asian Monsoon

AR5 reported *low confidence* in the observed weakening of the East Asian Monsoon (EAsiaM) since the mid-20th century. Since AR5, there has been improved understanding of changes in the EAsiaM, based on paleoclimatic evidence, instrumental observations and climate modeling simulations. Rainfall reconstructions from the Loess Plateau in China indicate that the northern extent of the monsoon rain belts migrated at least 300 km to the northwest from the LGM to the mid-Holocene (Yang et al., 2015). Similarly, Pliocene reconstructions indicate stronger intensity of the EAsiaM with a more northward penetration of the monsoon rain belt (Yang et al., 2018b). EAsiaM variability has been related to AMOC dynamics, especially during the last glacial period, but whether the relationship is negative or positive remains uncertain (Sun et al., 2012; Cheung et al., 2018; Kang et al., 2018).

Long-term precipitation observations from China indicate a trend of drying in the north and wetting in the central-eastern China along the Yangtze river valley since the 1950s (Qian and Zhou, 2014; Zhou et al., 2017a; Day et al., 2018), with a weakened EAsiaM low-level circulation that penetrates less far into northern China, increased surface pressure over northeast China and southward shift of the jet stream (Song et al., 2014). The southward shift and enhancement of the jet stream explains the increase of rainfall especially from the Meiyu front (Day et al., 2018) at the expense of drying over northeast China.

Anthropogenic factors such as GHGs and aerosols had an influence on the EAsiaM changes (Wang et al., 2013b; Song et al., 2014; Xie et al., 2016; Chen and Sun, 2017; Ma et al., 2017; Zhang et al., 2017a; Day et al., 2018; Tian et al., 2018) (Figure 8.11). Increased precipitation in the southern region has been linked to increased moisture flux convergence driven by GHG forcing whilst changes in anthropogenic aerosols have weakened the EAsiaM and reduced precipitation in the northern regions (Tian et al., 2018). Aerosol-induced cooling, associated atmospheric circulation changes and SST feedbacks weaken the EAsiaM and favour the observed dry-north and wet-south pattern of rainfall anomalies (Wang et al., 2013b; Song et al., 2014; Zhang et al., 2017a; Chen et al., 2018a, 2018b; Undorf et al., 2018).

Internal variability and volcanic eruptions also contributed to the weakened EAsiaM (Hsu et al., 2014; Qian and Zhou, 2014; Zhou et al., 2017b; Knutson and Zeng et al., 2018). Since the late 1970s, the EAsiaM weakening has been also linked to SST changes in the Pacific Ocean with warm conditions in the central-eastern tropical part and cold ones in the north, similar to a positive phase of the Pacific Decadal Variability (PDV, Section AVI.2.6) (Li et al., 2016c; Zhou et al., 2017b). In the late 1990s the transition from a positive to a negative PDV has been associated to the recent recovery observed in the EAsiaM strength (Zhou et al., 2017b). Atlantic Multidecadal Variability (AMV) also has an influence on the EAsiaM via the global teleconnection pattern propagating from the North Atlantic through the westerly jet (Zuo et al., 2013; Wu et al., 2016a, 2016b). This North Atlantic influence has contributed to the increase of precipitation over the Huaihe-Huanghe valley since the late 1990s (Li et al., 2017c). When PDV and AMV are in opposite phase, the former has a larger influence in driving the southern flooding and northern drought pattern over the region (Yang et al., 2017b).

In summary, there is strong evidence of a stronger EAsiaM and northward migration of the rainbelt during warmer climates based on paleoclimate reconstructions. There is *high confidence* that anthropogenic forcing has been influencing historical EAsiaM changes with drying in the north and wetting in the south observed since the 1950s, but there is *low confidence* in the magnitude of the anthropogenic influence. The transition towards a positive Pacific Decadal Variability phase has been one of the main drivers of the EAsiaM weakening since the 1970s *(high confidence)*.

8.3.2.4.3 West African Monsoon

Since AR5, there has been improved understanding of the West African Monsoon (WAfriM) response to natural and anthropogenic forcing. On paleoclimate timescales, enhanced summer insolation in the Northern Hemisphere intensified the WAfriM precipitation during the early-to-mid Holocene (*high confidence*), as seen in rainfall proxy records and climate model simulations (Masson-Delmotte et al., 2013; Mohtadi et al., 2016; Braconnot et al., 2019). Despite improvements in model simulations of the present-day monsoons, CMIP5 and CMIP6 models underestimate mid-Holocene changes in the amount and spatial extent of the WAfriM precipitation (Brierley et al., 2020) (Section 3.3.3.2).

During the recent past, long-term rain gauge observations display substantial variability in the WAfriM precipitation over the 20th century (Section 10.4.2.1). The WAfriM experienced the wettest decade of the 20th century during the 1950s and early 1960s (*high confidence*), over much of the western and central Sahel region, followed abruptly by the driest years during 1970-1989 (Ali and Lebel et al., 2009; Nicholson, 2013; Descroix et al., 2015). The percentage deficit in the annual rainfall during 1970-1989, relative to the long-term mean, ranged from 60% in the north of Sahel to 25-30% in the south (Le Barbé et al., 2002; Lebel et al., 2003). The long decline in annual rainfall is related to a decrease of rain occurrence over the Sahel (Le Barbé and Lebel, 1997; Frappart et al., 2009; Bodian et al., 2016) and the Soudano-Guinean sub-region of West-Africa (Le Barbé et al., 2002), even though the interannual variability pattern is more complex (Balme et al., 2006). Decrease of rainfall occurrences resulted from decreases in large convective events in the core of the rainy season (Bell et al., 2006), that modulate interannual variability of the WAfriM (Panthou et al., 2018).

Wetter conditions of the WAfriM prevailed later from the mid-to-late 1990s, although the positive trend in precipitation started since late 1980s (see also Section 10.4.2.1) over the Sahel (*high confidence*) and in the Guinean coastal region (*medium confidence*), indicating the geographical variation in the wetting recovery (Descroix et al., 2015; Sanogo et al., 2015; Bodian et al., 2016; Nicholson et al., 2018). While the interannual and decadal variability of annual rainfall is not homogeneous over the entire Sahel, the rainfall recovery was stronger in the east than in the west of the region (Nicholson et al., 2018) (Section 10.4.2.1). A shift in the seasonality of the Sahelian rainfall, including delayed cessation has also been reported (Nicholson, 2013; Dunning et al., 2018) (see also Section 10.4.2.1).

In the Sahel region, the emergence of this new rainfall regime is reflected in increased number of heavy and extreme events, compared to the 1970s-1980s, still not exceeding the values registered in the 1950s-1960s (Descroix et al., 2013, 2015, Panthou et al., 2014, 2018; Sanogo et al., 2015), and in higher interannual variability (Zhang et al., 2017b; Akinsanola and Zhou, 2020) associated with SST variations in the tropical Atlantic, Pacific and Mediterranean Sea (Rodríguez-Fonseca et al., 2015; Diakhate et al, 2019). Increased frequency of extreme rainfall events impacts high flow occurrences of the large Sahelian rivers as well as small to meso-scale catchments (Wilcox et al., 2018a). Overall, extreme intense precipitation events are more frequent in the Sahel since the beginning of the 21st century (Giannini et al., 2013; Panthou et al., 2014, 2018; Sanogo et al., 2015; Taylor et al., 2017). Intensification of mesoscale convective systems associated with extreme rainfall in the WAfriM is favoured by enhancement of meridional temperature gradient by the warming of the Sahara desert (Taylor et al., 2017) at a pace that is 2–4 times greater than that of the tropical-mean temperature (Cook et al., 2015c; Vizy et al., 2017). Periods of monsoon-breaks and the persistence of low rainfall events are still prominent, particularly after the onset, thus exposing West Africa simultaneously to the potential impacts of dry spells (Zhang et al., 2017b) and also extreme localized rains and floods (Engel et al., 2017; Lafore et al., 2017). Occurrence of extreme events is compounded by land use and land cover changes leading to increased runoff (Bamba et al., 2015; Descroix et al., 2018).

The Sahel drought from the 1970s until the early 1990s was related to anthropogenic emissions of sulphate aerosols in the Atlantic, which led to an inter-hemispheric pattern of SST anomalies and associated regional precipitation changes (Box 8.1; Section 6.3.3.2). Also the combined effects of anthropogenic aerosols and GHG forcing appear to have contributed to the late twentieth century drying of the Sahel through their effect on SST, by cooling the North Atlantic and warming the tropical oceans (Giannini and Kaplan, 2019; Hirasawa et al., 2020). Subsequent aerosol removal led to SST warming of the North Atlantic, shifting the ITCZ further northward and strengthening the WAfriM (Giannini and Kaplan, 2019). The recent recovery has been ascribed to prevailing positive SST anomalies in the tropical North Atlantic potentially associated with a positive phase of the Atlantic multidecadal oscillation (Diatta and Fink, 2014; Rodríguez-Fonseca et al., 2015). The Sahel rainfall recovery has also been attributed to higher levels of GHG in the atmosphere and increases in atmospheric temperature (Dong and Sutton, 2015).

In summary, most regions of West Africa experienced a wet period in the mid-20th century followed by a very dry period in the 1970s and 1980s that is attributed to aerosol cooling of the Northern Hemisphere (*high confidence*). Recent estimates provide evidence of a WAfriM recovery from the mid-to-late 1990s, with more intense extreme events partly due to the combined effects of increasing GHG and decreasing anthropogenic aerosols over Europe and North America (*high confidence*). On paleoclimate timescales, there is *high confidence* that the WAfriM strengthened during the early-to-mid Holocene in response to orbitally-forced enhancement of summer warming in the Northern Hemisphere.

8.3.2.4.4 North American Monsoon

Since AR5, there have been updates on the observed long-term variations and changes in the North American monsoon (NAmerM). During the Last Glacial Maximum (21,000-19,000 years ago), the NAmerM was substantially weaker due to cold, dry mid-latitude air associated with the Laurentide Ice Sheet (Bhattacharya et al., 2017b, 2018). The NAmerM strengthened until the mid-Holocene period, in response to ice sheet retreat and rising summer insolation, but probably did not exceed the strength of the modern system (*low confidence*), as indicated by model simulations (Metcalfe et al., 2015) and paleoclimatic reconstructions (Bhattacharya et al., 2018). Paleoclimatic evidence from proxy datasets and mid-Pliocene (PlioMIP1) simulations suggest a wetter southwestern United States during that warmer period (Haywood et al., 2013a; Pound et al., 2014; Ibarra et al., 2018) but it is not clear whether this is due to increases of precipitation associated with the monsoon or occurring during the winter season.

During 1948-2010, trends of boreal summer precipitation amount were significantly positive over New Mexico and the core NAmerM region, but significantly negative over southwestern Mexico (Hoell et al., 2016). In addition, diverse datasets like CRU, CHIRPS and GPCP show significant decreases of precipitation in parts of the southwest United States and northwestern Mexico, including the NAmerM region (Ashfaq et al., 2020; Cavazos et al., 2020). Other studies suggest a strengthening of the NAmerM upper level anticyclone since the mid-1970s, with a more frequent northward location (Diem et al., 2013). Between 1910-2010, the number of precipitation events increased across the northern Chihuahuan desert, within the NAmerM domain, despite a decrease in their magnitude, and the length of extreme dry and wet periods also increased (Petrie et al., 2014).

An increase in intense rainfall and severe weather events has been observed in several locations, especially in southwestern Arizona since 1991, resulting from increases in atmospheric moisture content and instability; a change that has been confirmed by convective-permitting model simulations (Luong et al., 2017; Pascale et al., 2019). A dense network of 59 rain gauges located in southeastern Arizona suggests an intensification of monsoon sub-daily rainfall since the mid-1970s (Demaria et al., 2019), as expected by a stronger global warming signature for sub-daily rather than daily or monthly precipitation accumulation (Section 11.4). Section 10.4.2.3 provides further details on changes in precipitation in southwestern North America. Evidence from multiple reanalyses suggests that increases in NAmerM rainfall have contributed to the increasing trend of global monsoon precipitation (Lin et al., 2014) (see also 2.3.1.4.2). In addition, more frequent occurrence of earlier retreats of the NAmerM since 1979 is documented (Arias et al., 2012, 2015), in association with the positive phase of the Atlantic Multidecadal Variability (AMV) and a westward expansion of the North Atlantic Subtropical High (Li et al., 2011, 2012b).

Analyses from a 50-km resolution GCM indicate that the NAmerM response to CO_2 is very sensitive to SST biases, showing reductions in summer NAmerM precipitation with increased CO_2 when the SST biases are small (Pascale et al., 2017) in contrast to CMIP5 models (Cook and Seager, 2013; Maloney et al., 2014; Torres-Alavez et al., 2014; Hoell et al., 2016). The NAmerM has been shown to be also sensitive to SO_2 emissions (García-Martínez et al., 2020).

In summary, both paleoclimate evidence and observations indicate an intensification of the NAmerM in a warmer climate (*medium confidence*). The intensification recorded since about the 1970s has been partly driven by GHG emissions (*medium confidence*).

8.3.2.4.5 South American Monsoon

Since AR5, there has been improved understanding of changes in the South American monsoon (SAmerM) as evidenced from paleoclimate records, instrumental observations and climate model simulations. However, GCMs still exhibit difficulties in reproducing SAmerM precipitation amount (Rojas et al., 2016; D'Agostino et al., 2020a). Paleoclimate evidence suggests a relatively stronger SAmerM during the 1400-1600 period (Bird et al., 2011b; Vuille et al., 2012; Ledru et al., 2013; Apaéstegui et al., 2014; Novello et al., 2016; Wortham et al., 2017). Last millennium GCM simulations are able to reproduce stronger SAmerM during the 1400-1600 period in comparison with warmer epochs such as the 900-1100 period (Rojas et al., 2016) or the current warming period (Díaz and Vera, 2018). PMIP3/CMIP5 simulations indicate a consistent weaker SAmerM during the mid-Holocene (6000 years ago; see Cross-Chapter Box 2.1) in comparison to current conditions (Bird et al., 2011a; Mollier-Vogel et al., 2013; Prado et al., 2013a; D'Agostino et al., 2020a), thus favouring savannah/grassland-like vegetation (Smith and Mayle, 2018), in agreement with climate reconstructions from different proxies (Prado et al., 2013b). Signals of weak and strong SAmerM during mid-Holocene and LGM, respectively, are evident also in high-resolution long-term (> ~ 22000 years) rainfall reconstructions based on oxygen isotopes in speleothems from Brazil (Novello et al., 2017; Stríkis et al., 2018b; Campos et al., 2019).

Isotope records from caves in the central Peruvian Andes show that the late Holocene (< 3000 years b.p) was characterized by multidecadal and centennial-scale periods of significant decline in intensity of the SAmerM

(Bird et al., 2011a; Vuille et al., 2012). This could be partly due to a reduction in the zonal SST gradient of the Pacific Ocean, favouring El Niño-like conditions (Kanner et al., 2013). Other studies suggest increased SAmerM precipitation amount during the Late Holocene, in association with the expansion of the tropical forest (Smith and Mayle, 2018). Well-dated equilibrium lines of glaciers during the deglaciation suggest that the AMOC enhances Atlantic moisture sources and precipitation amount increase over the tropical and southern Andes (Beniston et al., 2018).

Observations during 1979-2014 suggest that poleward shifts in the South Atlantic Convergence Zone (SACZ) noted in recent decades (Talento and Barreiro, 2018; Zilli et al., 2019), are associated with precipitation amount decrease along the equatorward margin and increase along the poleward margin of the convergenze zone (Zilli et al., 2019). Several observational studies identified delayed onsets of the SAmerM after 1978 related to longer dry seasons in the southern Amazon (Fu et al., 2013; Yin et al., 2014; Arias et al., 2015; Debortoli et al., 2015; Arvor et al., 2017; Giráldez et al., 2020; Haghtalab et al., 2020; Correa et al., 2021). In contrast, other studies indicate a trend toward earlier onsets of the SAmerM (Jones and Carvalho, 2013). These discrepancies are explained by the methodology used and the domain considered for the SAmerM, confirming the occurrence of delayed onsets of the SAmerM since 1978 (Correa et al., 2021). CMIP5 simulations show trends toward delayed onsets of the SAmerM in association with anthropogenic forcing, although the simulated trends underestimate the observed trends (Fu et al., 2013). Total rainfall reductions are observed in the southern Amazon during September-October-November after 1978 (Fu et al., 2013; Bonini et al., 2014; Debortoli et al., 2015, 2016; Espinoza et al., 2019), consistent with reductions in river discharge in the region (Molina-Carpio et al., 2017; Espinoza et al., 2019; Heerspink et al., 2020).

Significant increases in precipitation have been observed over southeastern Brazil during 1902-2005 while non-significant decreases have been found over central Brazil (Vera and Díaz, 2015). In Bolivia, increases were observed during 1965-1984, while reductions have occurred since then (Seiler et al., 2013). However, the Peruvian Amazon does not reveal significant changes in mean rainfall during 1965–2007 (Lavado et al., 2013; Ronchail et al., 2018). Historical simulations from CMIP5 ensembles adequately capture the observed summer precipitation amount over central and southeastern Brazil, thereby providing *high confidence* in interpreting the observed variability of SAmerM for the period 1960-1999 (Gulizia and Camilloni, 2015; Pascale et al., 2019). Also, CMIP5 simulations indicate that the anthropogenic forcing associated with increased GHG emissions is necessary to explain the positive trends in upper-troposphere zonal winds observed over the South American Altiplano (Vera et al., 2019). However, the detection of anthropogenically-induced signals for precipitation is still ambiguous in monsoon regions, like the SAmerM (Hoegh-Guldberg et al., 2019).

In summary, there is *high confidence* that the SAmerM onset has been delayed since the late 1970s. This is reproduced by CMIP5 simulations that consider anthropogenic forcing. There is also *high confidence* that precipitation during the dry-to-wet transition season has been reduced over the southern Amazon. Paleoclimate reconstructions and simulations suggest a weaker SAmerM during warmer epochs such as the Mid-Holocene or the 900-1100 period, and stronger monsoon during colder epochs such as the LGM or the 1400-1600 period (*high confidence*).

8.3.2.4.6 Australian and Maritime Continent Monsoon

Since AR5, several studies have examined observed variability and changes in the Australian and Maritime Continent Monsoon (AusMCM) using paleoclimate records, instrumental observations and modeling studies (Denniston et al., 2016; Zhang and Moise, 2016). Paleoclimate reconstructions and modelling indicate that the Indo–Australian monsoon may vary in or out of phase with the EAsiaM, depending on whether there is a meridional displacement or expansion of the tropical rainfall belt (Ayliffe et al., 2013; Denniston et al., 2016). For instance, mid-Holocene simulations suggest that the AusMCM weakens and contracts due to a decreased net energy input and a weaker dynamic component (D'Agostino et al., 2020a).

Rainfall increases have been observed over northern Australia since the 1950s, with most of the increases occurring in the north-west (Dey et al., 2018, 2019b; Dai, 2021) and decreases observed in the north-east (Li et al., 2012a) since the 1970s. There is also a trend towards more intense convective rainfall from thunderstorms over northern Australia (Dowdy, 2020). There is no consensus on the cause of the observed Australian monsoon rainfall trends, with some studies suggesting changes are due to altered circulation driving increased moisture transport or increased frequency of the wettest synoptic regimes (Catto et al., 2012; Clark et al., 2018). Other studies find that model simulations that include anthopogenic aerosols (Rotstayn et al., 2012; Dey et al., 2018) are better able to capture observed Australian monsoon rainfall trends than simulations with natural or GHG forcing only (Knutson and Zeng et al., 2018).

The Maritime Continent (MC) experiences the influence of both the Asian and the Australian monsoons, with rainfall peaking during boreal winter/austral summer (Robertson et al., 2011). Reductions in land rainfall and marine cloudiness over the MC and weakening of surface moisture flux convergence have been observed in the period 1950-1999 (Tokinaga et al., 2012; Yoden et al., 2017). These trends are indicative of a slowdown of the Walker Circulation, with positive sea level pressure trends over the MC and negative trends over the central equatorial Pacific (Tokinaga et al., 2012). More recently (1981-2014), a trend of increasing annual rainfall over large areas of the MC has been identified (Hassim and Timbal, 2019). Given the large variability in MC rainfall on interannual time scales, the choice of time period may influence the calculated rainfall trend (Hassim and Timbal, 2019).

During 1951-2007 daily rainfall extremes did not increase over the MC, in contrast to the rest of Southeast Asia (Villafuerte et al., 2015) (see Section 11.4.2). Rainfall extremes in Indonesia increased in austral summer, as evidenced from station weather observations for the period 1983-2012 (Supari et al., 2018).

In summary, notable rainfall increases have been observed in parts of northern Australia since the 1970s, although there is *low confidence* in the human contribution to these changes. Rainfall changes have been observed over the MC region but there is *low confidence* in the identification of trends because of large variability at interannual timescales.

8.3.2.5 Tropical cyclones

AR5 assessed *low confidence* in centennial changes in tropical cyclone activity globally, and in the attribution of observed changes in TCs to anthropogenic forcing. Since AR5, there has been considerable progress in understanding the observed changes of TCs and an overall improved knowledge of the sensitivity of TCs to both GHG and aerosol forcing (Knutson et al., 2019; Sobel et al., 2019).

There is *medium confidence* that anthropogenic forcing has contributed to observed heavy rainfall events over the United States associated with TCs (Kunkel et al., 2012) and other regions with sufficient data coverage (Bindoff et al., 2013) (see also Section 11.7.1.2). There has been increased frequency of TC heavy-rainfall events over several areas in the United States since the late 19th century that is greater than what would be expected solely from changes in U.S. landfall frequency, suggesting the increasing role of TCs have in causing heavy-rainfall events (Kunkel et al., 2010). For example, there is evidence for an anthropogenic contribution to the extreme rainfall of Hurricane Harvey in 2017 (Emanuel, 2017; Risser and Wehner, 2017; Van Oldenborgh et al., 2017; Trenberth et al., 2018; Wang et al., 2018c).

While TCs cause extreme local rainfall and flooding, they can be also an important contributor to annual precipitation and regional fresh water resources (Hristova-Veleva et al., 2020). Transport of moisture by TCs is an important contributor for precipitation over the coastal areas of East Asia mostly from July through October, with the TC rainfall accounting for nearly 10% to 30% of the total rainfall in the region (Guo et al., 2017b). Local TC rainfall totals depend on rain-rate and translation speed (the speed of TC movement along the storm track) with slow TCs such as Hurricane Harvey (2017), providing a clear example of the effect of slow translation speed on local rainfall accumulation, with urbanization exacerbating the storm total rainfall and flooding (Zhang et al., 2018d) (see Section 11.7.1).

In addition to evidence that rain-rates have increased, there is evidence that TC translation speed has slowed globally (Kossin, 2018) thus amplifying thermodynamic intensification of rainfall and may be linked to anthropogenic forcing (Gutmann et al., 2018). This is *limited evidence* however, so there is *medium confidence* of a detectable change in TC translation speed over the US. Since the 1900s, and there is *low confidence* for a global signal because of *limited agreement* among models and due to data heterogeneity. However, the slowdown is consistent with theoretical and modelling studies that indicate a general weakening of the tropical circulation with warming that reduces the speed of the TC system (Chauvin et al., 2017), though there is *limited* observational *evidence* (Sections 8.2.3.5, 11.7.1). Despite growing evidence that TC rainfall measures indicating increases, in general there is *low confidence* that a global anthropogenically forced trend in TC precipitation has been detected (Knutson et al., 2019), partly due to observational data limitations (e.g., Lau and Zhou, 2012).

In summary, there is *low confidence* of an observed increase in TC precipitation intensity due to observing system limitations. However, robust physical understanding (Section 8.2.3.2) and detailed singular event attribution studies provide evidence that tropical cyclone rainfall has increased with a warming climate (*high confidence*, Section 11.7.1.4).

8.3.2.6 Stationary waves

Stationary waves are planetary-scale waves that are approximately stable (stationary) in terms of geographic position, as opposed to propogating planetary waves, and are important both as part of the climatological general circulation and seasonal and shorter-term anomalies. They are related to surface features including land-ocean contrasts and major mountain ranges, as well as atmospheric features including the jet stream, storm tracks, and blocking, which are considered separately in the following sections. While zonal mean changes in P-E are dominated by thermodynamic effects (Section 8.2.2.1), changes in stationary waves are of key importance in understanding zonal asymmetries in the water cycle response to global warming (Wills and Schneider, 2015; Wills et al., 2019). AR5 did not explicitly assess stationary waves, but noted changes in related circulation features such as a *likely* poleward shift of the Northern Hemisphere storm tracks and an increase in frequency and eastward shift in North Atlantic blocking anticyclones, although there was *low confidence* in the global assessment of blocking.

Since AR5, several studies have demonstrated a link between stationary wave amplitude and wet and dry extremes in several different regions of the Northern Hemisphere (Liu et al., 2012; Coumou et al., 2014; Screen and Simmonds, 2014; Yuan et al., 2015) with changes in moisture transport playing an important role (Yuan et al., 2015). A 'resonance mechanism' has been proposed for an increasing amplitude of stationary waves (Petoukhov et al., 2013, 2016; Coumou et al., 2014; Kornhuber et al., 2017) and several studies have linked increasing amplitude of stationary waves to Arctic warming (Francis and Vavrus, 2012, 2015; Liu et al., 2012; Tang et al., 2014) as well as to global warming (Mann et al., 2017). However, other studies have not identified an increase in stationary wave amplitude (Barnes, 2013; Screen and Simmonds, 2013b, 2013a)

There has been considerable work on linkages (teleconnections) between Arctic warming and the mid-latitude circulation (see also Cross-Chapter Box 10.1). The limited amount of research on Southern Hemisphere stationary waves suggests changes in high latitude, mid-tropospheric stationary waves which influence Antarctic precipitation (Turner et al., 2017) and changes in stratospheric stationary waves that are associated with ozone depletion rather than increases in GHGs (Wang et al., 2013a). The observed climatology of Northern Hemisphere winter stationary waves is well-represented in the CMIP5 multi-model mean (Wills et al., 2019) but individual models have important deficiencies in reproducing stationary wave variability (Lee and Black, 2013). In the Southern Hemipshere, the observed climatology of stationary waves in CMIP5 models has considerable bias in both phase and amplitude (Garfinkel et al., 2020). A comprehensive assessment is not yet available for CMIP6 models.

In summary, there is *low confidence* in strengthened wintertime stationary wave activity over the North Atlantic, associated with increased poleward moisture fluxes east of North America There is *medium confidence* in a recent amplification of the Northern Hemisphere stationary waves in summer, but no formal attribution to anthropogenic climate change.

8.3.2.7 Atmospheric blocking

Atmospheric blocking refers to persistent, semi-stationary weather patterns characterized by a high-pressure (anticyclonic) anomaly that interrupts the westerly flow in the mid-latitudes of both hemispheres. By redirecting the pathways of mid-latitude cyclones, blocking can affect the water cycle and lead to negative precipitation anomalies in the region of the blocking anticyclone and positive anomalies in the surrounding areas (Sousa et al., 2017). In this way, blocking can also be associated with extreme events such as heavy precipitation (Lenggenhager et al., 2018), drought (Schubert et al., 2014) and heatwaves (Miralles et al., 2014a). AR5 reported *low confidence* in global-scale changes in blocking, due to methodological differences between studies.

Currently no consensus exists on observed trends in blocking during 1979-2013. (Horton et al., 2015) identified increasing trends in anticyclonic circulation regimes based on geopotential height fields in the mid troposphere, which may be partly related to the tropospheric warming itself and thus not represent real changes in the statistics of weather (Horton et al., 2015; Woollings et al., 2018). Hanna et al. (2018) and (Davini and D'Andrea, 2020) reported a significant increase in the frequency of summertime blocking over Greenland. A weakening of the zonal wind, eddy kinetic energy and amplitude of Rossby waves in summer in the Northern Hemisphere (Coumou et al., 2015, Kornhuber et al., 2017, 2019) and an increased 'waviness' of the jet stream associated with Arctic warming (Francis and Vavrus, 2015; Pfahl et al., 2015; Luo et al., 2019) have also been identified, which may be linked to increased blocking.

In contrast, it has been shown that observed trends in blocking are sensitive the choice of the blocking index, and that there is a large internal variability that complicates the detection of forced trends (Barnes et al., 2014; Cattiaux et al., 2016; Woollings et al., 2018), compromising the attribution of any observed changes in blocking. Many climate models still underestimate the occurrence of blocking, at least in winter over northeastern Atlantic and Europe (Dunn-Sigouin and Son, 2013), which leads to caution in the interpretation of their results for these regions. However, over the Pacific Ocean there have been large improvements in the simulation of blocking for the last 20 years (Davini & D'Andrea, 2016; Patterson et al., 2019). In the Southern Hemisphere, increases in blocking frequency have occurred in the South Atlantic in austral summer (Dennison et al., 2016) and in the southern Indian Ocean in austral spring (Schemm, 2018). A reduced blocking frequency has been found over the southwestern Pacific in austral spring (Schemm, 2018) (see also sections 2.3.1.4.3 and 3.4.1.3.3).

In summary, no robust trend in atmospheric blocking has been detected in modern reanalyses and in CMIP6 historical simulations (*medium confidence*). The lack of trend is explained by strong internal variability and/or the competing effects of low-level Arctic amplification and upper-level tropical amplification of the equator-to-pole temperature gradient (*medium confidence*).

8.3.2.8 Extratropical cyclones, storm tracks and atmospheric rivers

8.3.2.8.1 Extratropical cyclones and storm tracks

AR5 indicated *low confidence* in long-term changes in the intensity of extratropical cyclones (ETC) over the 20th century derived from centennial reanalyses and storminess proxies based upon sea level pressure. This was confirmed by the SREX assessment that the main Northern Hemisphere and Southern Hemisphere extratropical storm-tracks *likely* experienced a poleward shift during the last 50 years (Seneviratne et al., 2012) with *low confidence*, and inconsistencies within reanalysis datasets remain.

Since AR5 there has been considerable progress in quantifying storm-track activity using multiple reanalysis products and different methodologies (Hodges et al., 2011; Neu et al., 2013; Tilinina et al., 2013; Wang et al., 2016b). Over the Northern Hemisphere increases in the total number of cyclones from 1979 show a large spread of trends across different estimates (Neu et al., 2013; Li et al., 2016b; Grieger et al., 2018) (see also Section 2.3.1.4.3) resulting in *low confidence* in any clear increase of in the total number of cyclones. However, starting from the early 1990s, most reanalyses show increases in the total cyclone number by about

2-5% per decade (Figure 8.12). Increasing trends in the total number of cyclones are dominated by the increase in the number of shallow and moderate cyclones (which are more dependent on the datasets and identification mentods used) than with decreasing number of deep cyclones since the early 1990s (Tilinina et al., 2013; Chang, 2018). In the Southern Hemisphere the variability of the total number of cyclones is characterized by strong interdecadal variability preventing a clear assessment of trends. However, in contrast to the Northern Hemisphere,there is a significant increasing trend in the number of deep cyclones (~10% over 1979-2018) in ERA5, ERA-Interim, JRA55 and MERRA, and in the CFSR dataset after 2000 (Reboita et al., 2015; Wang et al., 2016b) (Figure 8.12).

Changes in the number of deep storms, which are often associated with heavier precipitiation over the North Atlantic and North Pacific, exhibit strong seasonal differences and decadal variability (Colle et al., 2015; Chang et al., 2016; Matthews et al., 2016, Priestley et al., 2020). An increase in the number of summer cyclones over the Atlantic-European sector (Tilinina et al., 2013) is consistent with the increase in the strength of the strongest fronts over Europe (Schemm et al., 2018). (Chang et al., 2016) reported a decrease in the number of strong summer storms in the latitudinal band 40º-75ºN over the last decades, however, the assessment of seasonal trends in the Atlantic-European sector is complicated by the choice of region, attribution of tracks to the region selected, and thresholds used to identify trajectories, leading to *low confidence* on regional seasonal trends. For the Southern Hemisphere, Grieger et al. (2018) reported growing number of cyclones over sub-Antarctic region in the austral-summer during 1979-2010, while statistically significant trends were absent during the austral winter.

[START FIGURE 8.12 HERE]

Figure 8.12: Annual anomalies (with respect to the reference period 1979-2018) of the total number of extratropical cyclones (a, c) and of the number of deep cyclones (<980hPa) (b, d) over the Northern (a, b) and the Southern (c, d) Hemispheres in different reanalyses (shown in colours in the legend). Note different vertical scales for panels (a), (b) and (c), (d). Thin lines indicate annual anomalies and bold lines indicate 5-yr running averages. (e), (f) The number of reanalyses (out of five) simultaneously indicating statistically significant (90% level) linear trends of the same sign during 1979–2018 for JFM over the Northern Hemisphere (e) and over the Southern Hemisphere (f). Updated from (Tilinina et al., 2013). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.12 HERE]

Analysis of storm-track activity over longer periods suffers from uncertainties associated with changing data assimilation and observations before and during the satellite era, resulting in inhomogeneities and discontinuities in centennial reanalyses (Krueger et al., 2013, Chang and Yau, 2016; Wang et al., 2013, 2016, Varino et al., 2019a). (Feser et al., 2015) reviewed multiple storm track records for the Atlantic-European sector and demonstrated growing storm activity north of 55°N from the 1970s to the mid 1990s with declining trend thereafter, sugesting strong interdecadal variability in storm track activity. This was also confirmed by (Krueger et al., 2019) from the analysis of geostrophic winds derived from sea level pressure gradients.

Poleward deflection of mostly oceanic winter storm tracks since 1979 was reported in both the North Atlantic and North Pacific (Tilinina et al., 2013; Wang et al., 2017b). This large-scale tendency has regional variations and may be seasonally dependent. Wise and Dannenberg (2017) reported a southward shift in the east Pacific storm-track from the 1950s to mid-1980s followed by northward deflection in the later decades. (King et al., 2019) resported an association of Atlantic storm track migrations with SSW events with Central and South European precipitation anomalies. Over centennial time-scales, Gan and Wu (2014) reported an intensification of stormtracks in the poleward and downstream regions of the North Pacific and North Atlantic upper troposphere using 20CR reanalysis. Poleward migration of the Southern Hemisphere storm-tracks (Grise et al., 2014; Wang et al., 2016b; Dowdy et al., 2019) was identified during the austral summer and is closely associated with cyclone-associated frontal activity (Solman and Orlanski, 2014, 2016) and cloud cover (Bender et al., 2012; Norris et al., 2016).

The representation of ETCs in both climate models and reanalyses is resolution-dependent, hence changes must be assessed with caution (Section 3.3.3.3). In particular, CMIP5 models show a systematic underestimation of the intensity of ETCs (Zappa et al., 2014), a feature that is partially related to their relatively coarse resolution or other possible deficiencies such as an excess of dissipation (Chang et al., 2013). The best representation of ETCs and their intensity in the North Atlantic are provided by relatively high horizontal resolution CMIP5 models (Zappa et al., 2014). Using a single high-resolution climate model, (Hawcroft et al., 2016) showed that precipitation amount associated with ETCs was generally well simulated though with too much precipitation during the strongest ECTs compared with observed estimations.

In summary, there is *low confidence* in recent changes in the total number of extra-tropical cyclones over both hemispheres. It is *as likely as not* that the number of deep cyclones over the Northern Hemisphere has decreased after 1979 and it is *likely* that the number of deep extra-tropical cyclones increased over the same period in the Southern Hemisphere. It is *likely* that extra-tropical cyclone activity in the Southern Hemisphere has intensified during austral summer with no significant changes in austral winter. There is *medium confidence* that boreal-winter storm tracks during the last decades experienced poleward shifts over the Northern and Southern Hemisphere oceans. There is *low confidence* of changes in extra-tropical cyclone activity prior 1979 due to inhomogeneities in the intrumental records and modern reanalyses.

8.3.2.8.2 Atmospheric rivers

Atmospheric rivers (ARs) are long, narrow (up to a few hundred km wide), shallow (up to few km deep) and transient corridors of strong horizontal water vapour transport that are typically associated with a low-level jet stream ahead of the cold front of an extratropical cyclone (Ralph et al., 2018). Atmospheric rviers were not assessed in AR5. ARs are associated with atmospheric moisture transport from the tropics to the mid and high latitudes (Zhu and Newell, 1998), although the drivers of moisture transport relative to the different airstreams within extratropical cyclones remains a subject of current study (Dacre et al., 2019). While much previous research has focused on the west coast of North America, ARs occur throughout extratropical and polar regions (e.g., Guan and Waliser, 2015) and are often associated with locally-heavy precipitation, including a substantial fraction of all midlatitude extreme precipitation events (e.g., Waliser and Guan, 2017). ARs also affect East Asia strongly during the period from late spring to summer (Kamae et al., 2017). ARs can be related to warming/melt events trough the intrusions of warm and moist air in Antarctica, Greenland and New Zealand (Bozkurt et al., 2018; Mattingly et al., 2018; Little et al., 2019), contributing about 45-60% of total annual precipitation in subtropical South America (Viale et al., 2018). They also transport moisture from South America to the western and central South Atlantic, feeding the ARs that reach the west coast of South Africa (Ramos et al., 2019). However, the estimation of precipitation rate from ARs can have large uncertainties, especially as ARs hit topographically complex coastal regions (Behrangi et al., 2016), which can cause complexities in quantifying AR-related precipitation.

Analysis of observed trends in the characteristics of ARs has been limited. (Gershunov et al., 2017) and (Sharma and Déry, 2019) have shown a rising trend in land-falling AR activity over the west coast of North American since 1948. (Gonzales et al., 2019) have also documented a seasonally-asymmetric warming of ARs affecting the United States West Coast since 1980, which has hydrological implications for the timing and magnitude of regional runoff. Longer-term paleoclimate analysis of ARs is even more limited, although Lora et al. (2017) reported that in the last glacial maximum, AR landfalls over the North America west coast were shifted southward compared to the present conditions.

In summary, it is *likely* that there was an increasing trend in the AR activity in the eastern North Pacific since the mid-20th century. However, there is *low confidence* in the magnitude of this trend and no formal attribution, although such an increase in activity is consistent with the expected and observed increase in precipitable water associated with human-induced global warming.

8.3.2.9 Modes of climate variability and regional teleconnections

Following on from the assessment in Chapters 2 and 3, this section considers changes in modes of variability at seasonal to interannual timescales in terms of their implications on recent water cycle changes. These modes are described in details in Technical Annex IV.

8.3.2.9.1 Tropical modes

The amplitude of the El Niño–Southern Oscillation (ENSO, Section AIV.2.3) variability has increased since 1950 (Section 2.4.2) but there is no clear evidence of human influence (Sections 2.4.2 and 3.7.3).

ENSO influences precipitation and evaporation dynamics, river flow and flooding at a global scale (see also Figure 3.37) (Ward et al., 2014, 2016; Martens et al., 2018). Reconstruction (1804-2005) of Thailand's Chao Praya River peak season streamflow displays a strong correlation with ENSO (Xu et al., 2019). Based on water storage estimates from 2002 to 2015, drought conditions over the Yangtze River basin followed La Niña events and flood conditions followed El Niño events (Zhang et al., 2015b). Strong correlation between ENSO and terrestrial water storage has been identified mostly in the subtropics but with diverse intensities and timelags depending on the region (Ni et al., 2018). The likelihood of increased/decreased flood hazard during ENSO events has a complex spatial pattern with large uncertainties (Emerton et al., 2017).

Tropical SSTs and associated global circulation may increase rainfall in West Africa, as observed in some years during 1950-2015, despite the presence of El Niño (Pomposi et al., 2020). During an El Niño summer, equatorial convective systems and the associated Walker circulation tend to shift eastward, leading to decreases in Indian summer monsoon rainfall (Li and Ting, 2015; Roy et al., 2019) (Lee and Tamas accepted). This teleconnection is modulated by Indian Ocean variability (Terray et al., 2021), as observed during the extreme positive IOD event in 2019 (Ratna et al., 2021). Since the end of the 19th century, synchronous hydroclimate changes (*medium confidence*) have been identified over south eastern Australia and South Africa (Gergis and Henley, 2017) modulated by ENSO, as well as other regional fluctuations like the Botswana High over Southern Africa (Driver and Reason, 2017). Over southern South America, the ENSO influence on precipitation (Cai et al., 2020; Poveda et al., 2020) interacts with the influence of SAM (Pedron et al., 2017), exhibiting large multi-decadal variations because of changes in the correlation between the two large-scale modes (Vera and Osman, 2018). Other processes underlying ENSO teleconnections of relevance for water cycle changes include water vapour and moisture transports, like over the Middle East (Sandeep and Ajayamohan, 2018), southeastern China (Yang et al., 2018c), or central Asia (Chen et al., 2018b), southeastern South America (Martín-Gómez and Barreiro, 2016; Martin-Gomez et al., 2016), Australia (Rathore et al., 2020) and southern United States (Okumura et al., 2017).

There is no evidence of trend in the Indian Ocean Dipole (IOD, Section AIV.2.4) mode and associated anthropogenic forcing (Sections 2.4.3 and 3.7.4). AR5 concluded that the IOD is *likely* to remain active, affecting climate extremes in Australia, Indonesia and east Africa. Since the AR5, IOD teleconnections have been identified extending further to the Middle East (Chandran et al., 2016), to the Yangtze river (Xiao et al., 2015), where in boreal summer and autumn positive IOD events tend to increase the precipitation in the southeastern and central part of the basin, and to the southern Africa extreme wet seasons (Hoell and Cheng, 2018). During the last millenium, the combined effect of a positive IOD and El Niño conditions have caused severe droughts over Australia (Abram et al., 2020). In the satellited period, it is found more effective in inducing significant decrease of rainfall over Indonesia, with the opposite occurring for negative IOD events (As-syakur et al., 2014; Nur'utami and Hidayat, 2016; Pan et al., 2018). Similarly, over the Ganges and Brahmaputra river basins major droughts have been recorded during co-occurring El Niño and positive IOD, while floods occurred during La Nina and negative IOD conditions (Pervez and Henebry, 2015). Over equatorial East Africa the IOD affects the short rain season (*medium confidence*) exacerbating flooding and inundations independently of ENSO (Behera et al., 2005; Conway et al., 2005; Ummenhofer et al., 2009; Hirons and Turner, 2018). Extreme conditions, like 2019 Australian bushfires and African flooding, have been associated with strong positive IOD conditions (Cai et al., 2021).

Intraseasonal variability, like the Madden Julian Oscillation (MJO, Section AIV.2.8) and the Boreal Summer Intraseasonal Oscillation (BSISO), are highly relevant to the water cycle (Maloney and Hartmann, 2000; Lee et al., 2013; Yoshida et al., 2014; Nakano et al., 2015). Since AR5, studies on MJO teleconnections within the tropics and from the tropics to higher latitudes have continued (Guan et al., 2012; Mundhenk et al., 2018; Tseng et al., 2019; Aberson and Kaplan, 2020; Finney et al., 2020b; Fowler and Pritchard, 2020; Fromang and Rivière, 2020).

The strength and frequency of the MJO have increased over the past century (*medium confidence*) (Oliver and Thompson, 2012; Maloney et al., 2019; Cui et al., 2020) because of global warming (Arnold et al., 2015; Carlson and Caballero, 2016; Wolding et al., 2017; Maloney et al., 2019). A 20th century reconstruction suggests a 13% increase of the MJO amplitude (Oliver and Thompson, 2012), with differences in seasonal variability (Tao et al., 2015; Wang et al., 2020e). However, up to half of changes recorded during the second half of the 20th century could be due to internal variability (Schubert et al., 2013). Other observed changes in MJO characteristics include a decrease (by 3-4 days) in the residence time over the Indian Ocean but an increase (by 5-6 days) over the Indo-Pacific and Maritime Continent sectors (Roxy et al., 2019).

Consequences of these changes are increased rainfall over Southeast Asia, northern Australia, southwest Africa and the Amazon, and drying over the west coast of the United States and Equador (Roxy et al., 2019). During the austral summer, air-sea interactions and location of the MJO active phase are important to modulate the strength of the rainfall response in the South Atlantic Convergence Zone (Shimizu and Ambrizzi, 2016; Alvarez et al., 2017), including its southward shift (Barreiro et al., 2019). In the austral winter, the intraseasonal variability is mostly influential over regions of the Amazon Basin (Mayta et al., 2019). Some MJO phases are particularly effective in conjuction with tropical cyclones in enhancing westerly moisture fluxes toward east Africa (Finney et al., 2020b).

Simulated changes in MJO precipitation amplitude are extremely sensitive to the pattern of SST warming (Takahashi et al., 2011; Maloney and Xie, 2013; Arnold et al., 2015) and ocean-atmosphere coupling (DeMott et al., 2019; Klingaman and Demott, 2020). In agreement with results from previous model generations, most CMIP5 models still underestimate MJO amplitude, and struggle to generate a coherent eastward propagation of precipitation and wind (Hung et al., 2013; Jiang et al., 2015; Ahn et al., 2017), affecting regional surface climate in the tropics and extra-tropics. In addition, most CMIP5 models simulate an MJO that propagates faster compared with observations, with a poorly represented intra-seasonal precipitation variability (Ahn et al., 2017). Over the Indian Ocean, the propagation speed of convection in some CMIP5 models tends to be slower than observed due to a strong persistence of equatorial precipitation (Hung et al., 2013; Jiang et al., 2015). Among other processes, improving the moisture-convection coupling, the representation of moist convection, the interaction between lower tropospheric heating and boundary layer convergence, and the topography of the Maritime Continent improve simulations of the MJO (Ahn et al., 2017, 2020a; Kim and Maloney, 2017; Yang and Wang, 2019; Tan et al., 2020a; Yang et al., 2020b). In fact, CMIP6 models reproduce the amplitude and propagation of the MJO better than CMIP5 models due to increased horizontal moisture advection over the Maritime Continent (Ahn et al., 2020b). Despite the diverse theories of MJO evolution and processes that have been developed since its discovery, a better understanding of its dynamics is still needed (Jiang et al., 2020; Zhang et al., 2020). Furthermore, metrics based on dynamical processes are needed to assess model simulations of these events (Stechmann and Hottovy, 2017; Wang et al., 2018a) as well as related teleconnections (Wang et al., 2020c).

In summary, multiple water cycle changes related to ENSO and IOD teleconnections have been observed across the 20th century (*high confidence*), mostly dominated by interannual to multi-decadal variations. The MJO amplitude has increased in the second half of the 20th century partly because of anthropogenic global warming (*medium confidence*) altering regional precipitation signals.

8.3.2.9.2 Extra-tropical modes

A positive trend has been observed in the Northern Annular Mode (NAM, Section AIV.2.1) in the second half of the 20th century, which partially reversed since the 1990s (Section 2.4.5.1), but the detection and attribution of these changes remain difficult (Section 3.7.1). The linkages of the NAM with weather and climate extremes in the northern extra-tropics are still unclear in models and observations (Vihma, 2014; Overland et al., 2016; Screen et al., 2018). However, robust links are identified between precipitation trends and variability in Europe and the phases of the Atlantic component of the NAM, i.e. the NAO (Moore et al., 2013; Comas-Bru and McDermott, 2014). Reduced winter precipitation is well correlated with the NAO over Southern Europe and Mediterranean countries (Kalimeris et al., 2017; Corona et al., 2018; Vazifehkhah and Kahya, 2018; Neves et al., 2019). NAO teleconnections in those regions include influences on groundwater and streamflow (Zamrane et al., 2016; Massei et al., 2017; Jemai et al., 2018). Remote teleconnections of the NAO have been identified over Northern China, the Yangtze River valley and India (Jin and Guan, 2017; Di Capua et al., 2020). The summer phase of the NAO is significantly correlated with variations in summer rainfall in East China, with the thermal forcing of the Tibetan Plateau providing a link to this Eurasian teleconnection (Wang et al., 2018f).

In the Southern Hemisphere, an observed positive trend is identified in the strength of the Southern Annular Mode (SAM, Section AIV.2.2) since 1950, especially in austral summer (*high confidence*, Section 2.4.1.2). While stratospheric ozone depletion and GHG increases largely contributed to this change, climate models still have trouble simulating the SAM and its response to ozone and GHGs (Section 3.7.2). Shifts in the southwesterly winds (Fletcher et al., 2018) and the expansion of the Southern Hemisphere Hadley cell (Kang and Polvani, 2011; Nguyen et al., 2018a) influence SAM-related rainfall anomalies in in southern South America and southern Australia during the austral spring–summer. Over New Zealand, large-scale SLP and zonal wind patterns associated with SAM phases modulate regional river flow (Li and McGregor, 2017). The SAM also influences precipitation and water vapour changes over Antarctica via moisture fluxes (Marshall et al., 2017; Oshima and Yamazaki, 2017; Grieger et al., 2018) but CMIP5 models are limted in their ability to simulate these regional teleconnections (Marshall and Bracegirdle, 2015; Palerme et al., 2017). SAM and its interaction with other large-scale modes of climate variability, like ENSO (Fogt et al., 2011) and the Indian Ocean Dipole (Hoell et al., 2017a), are responsible for fluctuations in Southern African rainfall (Nash, 2017) and southern South America (Gergis and Henley, 2017). In May, the SAM can trigger a southern Indian Ocean dipole SSTA favoring more or less precipitation over the Indian sub-continent and adjacent areas (Dou et al., 2017), also affecting subsequent summer monsoon in the South China Sea (Liu et al., 2018d). Over South America, a positive SAM is associated with dry conditions (Holz et al., 2017) due to reduced frontal and orographic precipitation and weakening of moisture convergence. Regions particularly affected include Chile (Boisier et al., 2018) and the rivers of central Patagonia (Rivera et al., 2018).

In summary, while the attribution of 20th century variations of the NAM/NAO is still unclear, there is a strong relationship with precipitation changes over Europe and in the Mediterranean region (*high confidence*). SAM teleconnections are associated with changes in moisture transport and extend to South America, Australia and Antarctica (*high confidence)* with documented drying occurring as a result of the *very likely* human-induced SAM trend toward its positive phase observed from the 1970s until the 1990s (Section 3.7.2).

8.4 What are the projected water cycle changes?

We consider global and regional climate projections of the water cycle, assessing projected changes in each component of the water cycle (Section 8.4.1) and the global-scale and regional phenomena that directly impact it (Section 8.4.2).

8.4.1 Projected water cycle changes

Most projected changes in the water cycle are not expected to be uniform in space or time. They are driven by both dynamical and thermodynamical processes (Section 8.2) and have not necessarily emerged yet in the recent observational record (Section 8.3) as they are superimposed on substantial natural fluctuations in weather and climate. Therefore, projecting regional water cycle changes remains challenging. However, a number of physically understood responses can be evaluated using both CMIP5 and CMIP6 models, which are important for guiding decision making that anticipates, prepares for, and responds to water cycle changes. In this section, global maps of projected changes in water cycle variables are assessed using the WGI AR6 'simple method' (see Cross-Chapter Box Atlas1), which uses hatching to highlight where less than 80% of the models agree on the sign of projected changes. This choice differs from Section 4.2.6 for a number of reasons. These include the weak signal-to-noise ratio of projected hydrological changes in low to medium emission scenarios, the sensitivity of their statistical significance to the baseline reference period, and the non-Gaussian distribution of many water cycle variables (see CCB Atlas.1 for more details on strengths and limitations of the hatching methods implemented within AR6).

8.4.1.1 Global water cycle intensity and P-E over land and oceans

As discussed in 8.3.1.1, the definition of global water cycle intensity varies from the simple metric of increases in global mean precipitation to broader joint considerations of water vapour and its transport, precipitation minus evaporation (P-E) rates and continental runoff (see Figure 8.1). AR5 determined that globally averaged precipitation is *virtually certain* to increase with temperature and that there is *high confidence* that the contrast of annual mean precipitation between dry and wet regions and seasons will increase over most of the globe as temperatures and moisture transports increase (Collins et al., 2013). AR5 also highlighted that continued ocean warming for a few decades after GHG forcing stabilizes or begins to decrease will also lead to further increases in global mean precipitation and evaporation.

In this Report, Chapter 4 provides an updated assessment of global annual precipitation (Section 4.3.1), finding that it is *very likely* that annual precipitation averaged over all land regions continuously increases as global surface temperatures increase in the 21st century (*high confidence*). CMIP6 projections for long-term changes in P-E (Figure 8.13) show that, for all scenarios, P-E increases over the tropics and high latitudes and decreases over the subtropics, resulting from a thermodynamically driven amplification of P-E patterns (Section 8.2.2.1). Both the intensity of changes and the spread among the models is larger for the higher emission scenarios. A less coherent latitudinal pattern and smaller magnitude of P-E changes over land reflect the complex influence of land-ocean warming contrast, atmospheric circulation change and vegetation feedbacks (Section 8.2.2.1). However, stronger atmospheric moisture transport, increases in precipitation and evaporation over global land and ocean and larger continental runoff that is in part fed by melting of glaciers characterises a more intense water cycle with global warming.

Global and global land mean water cycle changes from CMIP6 projections are shown in Table 8.1. Increases in global and continental precipitation, P-E and runoff in both the mid-term and long-term illustrate the future intensification of the water cycle, with the magnitude of change increasing with emission scenario. Consistent with AR5, CMIP6 simulations of global mean precipitation show a systematic multi-model mean increase of 1.6% to 2.9 % per 1°C warming (apparent hydrological sensitivity; Section 8.2.1) by 2081-2100 relative to present day across the new SSP scenarios (using global surface air temperature change from Table 4.1). It is well understood that rising concentrations of CO_2 drive a long-term increase in global precipitation with warming, but with the increase partly offset by rapid atmospheric adjustments to the direct atmospheric heating from radiative forcing agents (Section 8.2.1). The largest apparent hydrological sensitivity is found for SSP1-1.9, where the suppressing effects on precipitation from atmospheric heating by greenhouse gases rapidly reduced as their concentration falls. Additional warming due to reduced aerosol loadings under the SSP scenarios (Lund et al., 2019) further increases global precipitation (Rotstayn et al., 2013; Wu et al., 2013; Salzmann, 2016; Richardson et al., 2018b; Samset et al., 2018b; Westervelt et al., 2018), with particularly strong contributions from increased monsoon rainfall over East and South Asia (Levy et al., 2013; Westervelt et al., 2015; Dwyer and O'Gorman, 2017).

Over global land there is a small range in global mean multi-model mean precipitation increase across scenarios in the mid-term (2.6-4.0%), which widens (to 2.6-8.8%) in the long-term (Table 8.1). The long term projections are consistent with the Chapter 4 assessment that global annual precipitation over land is projected to increase on average by 2.4 [-0.2 to 4.7] % (*very likely* range) in the SSP1-1.9 low-emission scenario and by 8.3 [0.9 to 12.9] % in the SSP5-8.5 high-emission scenario by 2081–2100 relative to 1995–2014. Small differences in assessed model mean changes in Chapter 4, Table 4.2 result from a slightly different set of models considered for Table 8.1. Over land, P-E increases by ~2-3% in the mid-term (apart from SSP5-8.5 where increases are almost 5%) and ~1-12% in the long-term, determined by increased moisture transport from the ocean to land (Section 8.4.1.2). Runoff increases are larger and less certain due to additional inputs from glacier melt and changes in groundwater storage (Section 8.4.1.7). Overall, precipitation and runoff are *very likely* to increase over the global land in all scenarios in the mid term and long term. P-E is *likely* to increase over global land in the mid and long term and *very likely* in SSP1-1.9, SSP3-7.0 and SSP5-8.5 pathways. The mid-term consistency in projections across scenarios is not apparent for precipitable water vapour, which increases over land by around 6-15% in the mid-term and 5-36% in the long-term across all scenarios. This implies that increases in extreme precipitation (closely related to atmospheric water vapour content; Section 8.2.3.2) are dependent on mitigation pathway, even in the mid-term (Section 11.4.5). Water vapour residence time (computed as the ratio of precipitable water vapour to precipitation from values in Table 8.1) increases from 8 days in the present to 9 days in mid-term and up to about 10 days in the long-term over land in SSP3-7.0, indicating a longer time to moisten the atmosphere between precipitation events. The CMIP6 projections are therefore consistent with an intensification but not acceleration of the global water cycle.

In summary, it is *virtually certain* that global water cycle intensity, considered in terms of global and continental mean precipitation, evaporation and runoff, will increase with continued global warming. Global annual precipitation over land is projected to increase on average by 2.4 [-0.2 to 4.7] % (*very likely* range) in the SSP1-1.9 low-emission scenario and by 8.3 [0.9 to 12.9] % in the SSP5-8.5 high-emission scenario by 2081–2100 relative to 1995–2014.

Table 8.1: Global and global land annual mean water cycle projections in the medium term (2041-2060) and long term (2081-2100) relative to present day (1995-2014), showing present day mean and 90% confidence range across CMIP6 models (historical experiment) and projected mean changes and the 90% confidence range across the same set of models and a range of shared socioeconomic scenarios. Note that the exact value of changes can vary slightly based on the number of models assessed, but not sufficiently to affect the assessment. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

		Mid term: 2041–2060 minus reference period					***Long term: 2081–2100 minus reference period***				
	1995-2014 reference period	**SSP1-1.9**	**SSP1-2.6**	**SSP2-4.5**	**SSP3-7.0**	**SSP5-8.5**	**SSP1-1.9**	**SSP1-2.6**	**SSP2-4.5**	**SSP3-7.0**	**SSP5-8.5**
Global Annual											
Precipitation (mm/day)	2.96 [2.76-3.17]	0.06 [0.03-0.11]	0.07 [0.03-0.12]	0.07 [0.04-0.12]	0.06 [0.03-0.11]	0.08 [0.03-0.14]	0.06 [0.02-0.11]	0.09 [0.04-0.17]	0.12 [0.07-0.21]	0.15 [0.08-0.24]	0.2 [0.1-0.33]
Precipitable Water (kg/m2)	24.79 [23.06-26.82]	1.42 [0.7-2.26]	1.84 [1.03-2.62]	2.29 [1.6-3.09]	2.7 [1.92-3.92]	3.15 [2.13-4.38]	1.11 [0.28-2.13]	2.11 [0.98-3.15]	3.76 [2.41-5.08]	6.2 [4.24-8.83]	7.92 [5.21-10.69]
Global Land Annual											
Precipitation (mm/day)	2.27 [1.98-2.58]	0.07 [0.02-0.11]	0.07 [-0.0-0.13]	0.06 [0.01-0.13]	0.06 [0.02-0.12]	0.09 [0.01-0.16]	0.06 [0.01-0.1]	0.08 [0.02-0.16]	0.11 [0.02-0.19]	0.14 [0.03-0.22]	0.2 [0.07-0.32]
Precipitation - Evaporation (mm/day)	0.87 [0.49-1.26]	0.02 [0.0-0.03]	0.02 [-0.01-0.05]	0.02 [-0.02-0.06]	0.03 [-0.0-0.06]	0.04 [0.0-0.1]	0.01 [-0.0-0.03]	0.03 [-0.01-0.08]	0.04 [-0.01-0.07]	0.07 [0.0-0.12]	0.1 [0.01-0.22]
Runoff (mm/day)	0.79 [0.54-1.0]	0.02 [-0.0-0.05]	0.04 [-0.0-0.1]	0.04 [-0.0-0.11]	0.04 [0.01-0.08]	0.06 [0.01-0.14]	0.02 [-0.0-0.03]	0.04 [-0.0-0.13]	0.06 [0.0-0.17]	0.1 [0.02-0.2]	0.15 [0.04-0.27]
Precipitable Water (kg/m2)	18.86 [17.12-21.28]	1.23 [0.57-1.96]	1.58 [0.77-2.42]	1.96 [1.34-2.76]	2.33 [1.63-3.46]	2.72 [1.79-3.84]	0.95 [0.19-1.95]	1.78 [0.8-2.77]	3.18 [2.04-4.34]	5.33 [3.57-7.5]	6.81 [4.35-9.32]

[START FIGURE 8.13 HERE]

Figure 8.13: Zonal and annual mean projected long-term changes in the atmospheric water budget. Zonal and annual mean projected changes (mm/day) in P (precipitation, left column), E (evaporation, middle column), and P-E (right column) over both land and ocean areas (thick line) and over land only (dashed line) averaged across 36 to 38 CMIP6 models in the SSP1-2.6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario, respectively. Shading denotes confidence intervals estimated from the CMIP6 ensemble under a normal distribution hypothesis. Colour shading denotes changes over both land and ocean. Grey shading represents internal variability derived from the pre-industrial control simulations. All changes are estimated for 2081-2100 relative to the 1995-2014 base period. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.13 HERE]

8.4.1.2 Water vapour and its transport

Globally, AR5 assessed that by the end of the 21st century, the average quantity of water vapour in the atmosphere could increase by 5 to 25%, depending on emissions. AR5 assessed that increases in near-surface specific humidity over land are *very likely*, but that it was also *likely* that near-surface relative humidity would decrease over many land areas, although with only *medium confidence*. In terms of moisture transport, AR5 assessed that it was *likely* that moisture transport into the high latitudes would increase and that there was *high confidence* that, over the ocean, atmospheric moisture transport from the evaporative regions to the wet regions would increase.

CMIP6 climate models continue to project a steady increase in global mean column-integrated water vapour by around 6-13% by 2041-2060 and 5-32% by 2081-2100, depending on scenario (Table 8.1). This is consistent with projected atmospheric warming (Section 4.5.1.2) and the Clausius-Clapeyron relationship (Section 8.2.1) where every degree Celsius of warming is associated with a ~7% increase in atmospheric moisture in the lower atmospheric layers where most of the water vapour is concentrated. This increase sustains a positive feedback on anthropogenic global warming (Section 7.4.2.2). In contrast, the response of clouds is much more spatially heterogeneous, microphysically complex, and model-dependent so that the projected cloud feedbacks remain a key uncertainty for constraining climate sensitivity (Section 7.4.2.4).

CMIP6 models project an overall decrease in near-surface relative humidity over land, although with some regional and seasonal variations in their response (Fig. 4.26). Regional changes in near-surface humidity over land are dominated by thermodynamic processes and are primarily controlled by moisture transport from the warming ocean (Chadwick et al., 2016b). Increases in specific humidity lower than the thermodynamic rate are explained by greater warming over land than ocean and modulated by land-atmosphere feedbacks such as soil moisture and plant stomatal changes (Section 8.2.2.1) (Berg et al., 2017; Douville et al., 2020). This explains why climate models continue to project a contrasting response of near-surface relative humidity, with a slight and possibly overestimated increase over the oceans and a consistent but possibly underestimated decrease over land (Byrne and O'Gorman, 2016; Douville and Plazzotta, 2017; Zhang et al., 2018c).

While projections of water vapour are well understood due to the constraints of the Clausius-Clapeyron relationship, projections of water vapour transport are complicated regionally by the role of changes in the wind field, which is influenced by a wide variety of factors. Additionally, there has been relatively little general evaluation of moisture transport in models. In CMIP5 models, both the mean and variability of the vertically-integrated moisture transport is projected to increase, largely due to increases in water vapour (Lavers et al., 2015), with substantial regional differences (Levang and Schmitt, 2015). Single model studies have illustrated projected increases in low-altitude moisture transport into convergence regions (Allan et al., 2014) and from ocean to land (Zahn and Allan, 2013) that are consistent with present day trends. Increases in moisture transport have been linked to increases in large precipitation accumulations over land (Norris et al., 2019). Based on robust physics and supported by modelling studies, it is well understood that moisture transport increases into convergent parts of the atmospheric circulation such as storm systems, the tropical rain belt and high latitudes (Section 8.2.2.1), but changes in atmospheric circulation that are less well understood alter moisture transport regionally (Section 8.2.2.2). Therefore, given the limited examination of moisture transport in models, regional projections should be considered with caution. Changes in moisture transport specifically associated with monsoons, atmospheric rivers, and other specific circulation features are discussed further in the following sections.

In summary, there is *high confidence* in continued increases in global mean column integrated water vapour and near-surface specific humidity over land. There is *medium confidence* in region and season-dependent decreases in near-surface relative humidity over land, due to the complex physical processes involved. In general, there will be increases in moisture transport into storm systems, monsoons and high latitudes (*medium confidence*).

8.4.1.3 Precipitation amount, frequency and intensity

This section assesses projected changes in precipitation at regional scales. Note that changes in precipitation seasonality are assessed in Box 8.2 and that changes in regional monsoons are assessed in Section 8.4.2.4, where both circulation and rainfall are considered. Further assessments of regional projections of precipitation are presented in Chapters 10, 12 and the Atlas, while a comprehensive assessment of changes in precipitation extremes is provided in Chapter 11.

AR5 assessed that the contrast of mean precipitation amount between dry and wet regions and seasons is expected to increase over most of the globe as temperatures increase (*high confidence*), but with large regional variations. Precipitation over the high latitudes, equatorial Pacific Ocean, mid-latitude wet regions, and monsoon regions were assessed as *likely* to increase under the RCP8.5 scenario, and in many mid-latitude and subtropical dry regions as *likely* to decrease (AR5 Chapters 7, 12, and 14). Extreme precipitation over most mid-latitude land areas and wet tropical regions was assessed as *very likely* to become more intense and more frequent.

Geographical patterns of projected precipitation changes show substantial seasonal contrasts and regional differences, including over land (Figure 8.14; Figure 4.27). Projections for 2081-2100 under the SSP2-4.5 scenario suggest increased precipitation over the tropical oceans, northeastern Africa, the Arabian Peninsula, India, southeastern Asia and the Polar Regions while decreased precipitation is projected mainly over the subtropical regions (Section 4.5.1.4). Precipitation changes contrast regionally in the tropics with wetter wet seasons over South Asia, central Sahel and eastern Africa, but less precipitation over Amazonia and coastal West Africa (Section 8.4.2.4). These large-scale responses are associated with stronger moisture transports in a warmer climate that are modulated by the greater warming over land than ocean, atmospheric circulation responses and land surface feedbacks (Section 8.2.2). There is agreement across CMIP5 and CMIP6 modelling studies that precipitation increases in wet parts of the atmospheric circulation and decreases in dry parts (Liu and Allan, 2013; Kumar et al., 2015; Deng et al., 2020; Schurer et al., 2020) although these regions shift with atmospheric circulation changes. The overall pattern is robust across different model scenarios and time horizons (Tebaldi and Knutti, 2018), but some deviations from the mean pattern cannot be excluded due to the multiple timescales and non-linear atmospheric or land surface processes involved (Section 8.5.3). Near-term regional changes in precipitation are more uncertain because of a stronger sensitivity to natural variability (Section 8.5.2) and non-GHG anthropogenic forcings (Section 4.4.1.3 and 8.4.3.1).

Projected changes in regional precipitation also arise as a response to changes in large-scale atmospheric circulation (Section 8.2.2.2 and 8.4.2), both in the tropics (Chadwick et al., 2016a; Byrne et al., 2018) and extratropics (Shaw, 2019; Oudar et al., 2020b). Despite variability in simulated changes, CMIP5 climate models consistently project large rainfall changes (of varying sign) over considerable proportions of tropical land during the 21st century (Chadwick et al., 2016a). Since AR5, some robust responses in large-scale circulation patterns have been identified. For example, and as further assessed in Section 8.4.2, CMIP6 models project a northward shift in the tropical rain belt over eastern Africa and the Indian Ocean and a southward shift in the eastern Pacific and Atlantic oceans (Mamalakis et al., 2021). A projected strengthening and tightening of the tropical rain belt increases the contrasts between wet and dry tropical weather regimes and seasons. It is less clear how the well understood poleward expansion of the subtropics and mid-latitude storm tracks influences precipitation over sub-tropical and mid-latitude continents (Section 8.2.2.2).

An ensemble of 31 CMIP6 models under the SSP5-8.5 scenario projects increases of precipitation by 10–30% over much of the United States and decreases by 10–40% over Central America and the Caribbean by 2080-2099 (Almazroui et al., 2021). This CMIP6 ensemble also projects an increase in annual precipitation over the southern Arabian Peninsula and a decrease over the northern Arabian Peninsula, as also projected by CMIP3 and CMIP5 models (Almazroui et al., 2020a). Annual mean precipitation is projected to increase over South Asia during the 21st century under all scenarios, although the rate of change varies within the region based on 27 CMIP6 models (Almazroui et al., 2020c). CMIP6 projections also display a reduction in annual mean precipitation over northern and southern Africa while increases are projected over central Africa, under the SSP1-2.6, SSP2-4.5 and SSP5-8.5 scenarios (Almazroui et al., 2020b). The AR6 Atlas assesses that regions where annual mean rainfall is *likely* to increase include the Ethiopian Highlands, East, South and North Asia, southeastern South America, northern Europe, northern and eastern North America, and the Polar Regions. In contrast, regions where annual mean rainfall is *likely* to decrease include southern Africa, coastal West Africa, Amazonia, southwestern Australia, Central America, southwestern South America, and the Mediterranean.

AR5 identified that high-latitude precipitation increase may lead to an increase in snowfall in the coldest regions and a decrease of snowfall in warmer regions due to a decreased number of freezing days. The fraction of precipitation falling as snow and the duration of snow cover was projected to decrease. Heavy snowfall events globally are not expected to decrease significantly with warming as they occur close to the water freezing point, which will migrate poleward and in altitude (O'Gorman, 2014; Turner et al., 2019). There are only a small number of studies evaluating the implications of this mechanism in specific regions. A study for the northeastern US indicates smaller reductions for major snowfall events against the broader decline in snowfall expected from thermodynamic effects (Bintanja and Andry, 2017). Arctic snowfall is projected to decrease as rainfall makes up more of the precipitation (Zarzycki, 2018).

Beyond annual or seasonal mean precipitation amounts, an implication of the parallel intensification of the global water cycle and of the increased residence time of atmospheric water vapour (Section 8.2.1) is that the distribution of daily and sub-daily precipitation intensities will experience significant changes (Pendergrass and Hartmann, 2014b; Pendergrass et al., 2015; Bador et al., 2018; Douville and John, 2020), with fewer but potentially stronger events (*high confidence,* Section 4.3.3). CMIP6 projections show that in the long-term more drier days but more intense single events of precipitation are expected, regardless of scenario (Figure 8.15). Over almost all land regions, it is *very likely* that extreme precipitation will intensify at a rate close to the 7% per 1°C of global warming, but with large spatial differences (Section 11.4; Section 8.2.3.2). The projected increase in precipitable water is expected to lead to an increase in the highest possible precipitation intensities and an increase in the probability of occurrence of extreme precipitation events on the global scale (Neelin et al., 2017), regardless of how annual mean precipitation changes (O'Gorman and Schneider, 2009; O'Gorman, 2015). The projected increase in heavy precipitation intensity is also found for daily mean precipitation intensity though at a lower rate (Pendergrass and Hartmann, 2014a).

An increase in the number of dry days is also projected in several regions of the world (Polade et al., 2014, (Polade et al., 2014; Berthou et al., 2019a)), which can dominate the annual precipitation change at least in the subtropics (Polade et al., 2014; Douville and John, 2020). These findings are supported by CMIP6 projections showing a widespread increase in daily mean precipitation intensity over land (Fig.8.15bdf) as well as an increase in the number of dry days in the subtropics and over Amazonia and Central America (Fig.8.15abc). Such changes in precipitation regimes, as well as the general increase in the frequency and intensity of precipitation extremes (Section 11.4.5), contribute to an overall increase in precipitation variability (Polade et al., 2014; Pendergrass et al., 2017; Douville and John, 2020). This is also found in

CMIP6 models, which show a stronger increase of interannual variability than in seasonal mean precipitation changes, apart from in the winter extratropics where both quantities increase at the same rate with increasing global warming levels (Fig.8.16).

In summary, it is *virtually certain* that global precipitation will increase with warming due to increases in GHG concentrations and decreases in air pollution. There is *high confidence* that total precipitation will increase in the high latitudes, with a shift from snowfall to rainfall except in the coldest regions and seasons. There is also *high confidence* that precipitation will decrease over the Mediterranean, southern Africa, Amazonia, Central America, southwestern South America, southwestern Australia and coastal West Africa and that monsoon precipitation will increase over South Asia, East Asia and central-eastern Sahel. See Section 8.4.2.4 for a more detailed assessment of changes in regional monsoons. Daily mean precipitation intensities, including extremes, are projected to increase over most regions (*high confidence*). The number of dry days is projected to increase over the subtropics, Amazonia, and Central America (*medium confidence*). There is *high confidence* in an overall increase in precipitation variability over most land areas.

[START FIGURE 8.14 HERE]

Figure 8.14: Projected long-term relative changes in seasonal mean precipitation. Global maps of projected relative changes (%) in seasonal mean of precipitation averaged across 29 CMIP6 models in the SSP2-4.5 scenario. All changes are estimated for 2081-2100 relative to the 1995-2014 base period. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.14 HERE]

[START BOX 8.2 HERE]

BOX 8.2: Changes in water cycle seasonality

Observed changes

AR5 did not highlight observed changes in water cycle seasonality and SRCCL mostly emphasized changes in vegetation seasonality. Since AR5, a number of relevant studies have been published, but often with conflicting results. Based on three *in situ* datasets, reduced precipitation seasonality was identified over 62% of the terrestrial ecosystems analysed from 1950-2009 (Murray-Tortarolo et al. 2017). In contrast, both *in situ* and satellite data show a general increase in the annual range of precipitation from 1979 to 2010, which is dominated by wetter wet seasons (Chou et al., 2013). This paradox may be partly explained by a larger aerosol radiative forcing in the middle of the 20th century as well as by internal variability (Kumar et al., 2015; see also Box 8.1). For instance, the "long rains" over East Africa experienced declining trends in the 1980s and 1990s (Nicholson, 2017), which was linked to anthropogenic aerosols and SST patterns (Rowell et al., 2015), followed by a recent recovery that was linked to internal variability (Wainwright et al., 2019). Two satellite datasets revealed decreased rainfall seasonality in the tropics but an increased seasonality in the subtropics and mid-latitudes since 1979, without clear attribution (Marvel et al., 2017).

Large differences have been found across seven global precipitation datasets, with no region showing a consistent, statistically significant, positive or negative trend over the last three decades (Tan et al., 2020b). Regional studies suggest that observed changes in precipitation seasonality are neither uniform nor stable across the 20th century (Li et al., 2016a; Mallakpour and Villarini, 2017; Sahany et al., 2018; Deng et al., 2019). Since the 1980s, there is however growing evidence that contrasts between wet and dry regimes, including seasonality, have increased (Liu and Allan, 2013; Polson et al., 2013; Murray-Tortarolo et al., 2016; Tapiador et al., 2016; Gallego et al., 2017; Polson and Hegerl, 2017; Barkhordarian et al., 2018; Lan et al., 2019; Liang et al., 2020; Schurer et al., 2020).

Additional changes in seasonality may manifest in the timing and duration of wet seasons. A later monsoon onset trend was reported throughout India from 1901 to 2013 (Sahany et al., 2018). Conversely, an earlier rainfall onset was implicated in increased springtime rainfall over the Tibetan Plateau in recent decades (Zhang et al., 2017c). Winter and early spring precipitation over the northwestern Himalaya for the period 1951-2007 shows an increasing trend of daily precipitation extremes in association with enhanced amplitude variations of extra-tropical synoptic-scale systems known as "Western Disturbances" (Cannon et al., 2015; Krishnan et al., 2019; Madhura et al., 2014). In China, an earlier onset was observed during 1961-2012 (Deng et al., 2019). In the African Sahel, rainfall has been most concentrated in the peak of the rainy season since the end of the 20th century (Biasutti, 2019). A shift in the seasonality of Sahelian rainfall, including delayed cessation has also been reported (Nicholson, 2013; Dunning et al., 2018) (Section 10.4.2.1). Over southern Africa, an observed earlier onset (1985-2007) is in contrast to a simulated historical and projected future delay in the wet season (Maidment et al., 2015; Dunning et al., 2018). An increasingly early onset of the North American monsoon has been observed from 1978 to 2009 (Arias et al., 2015). Seasonality changes in the South American monsoon indicate delayed onsets since 1978 (Fu et al., 2013; Yin et al., 2014; Arias et al., 2015; Debortoli et al., 2015; Arvor et al., 2017; Giráldez et al., 2020; Haghtalab et al., 2020; Correa et al., 2021).

In northern high latitudes, a shorter snow season (Zeng et al., 2018a) is mainly due to an earlier onset of spring snowmelt (Peng et al., 2013) which has been attributed to anthropogenic climate change (Najafi et al., 2016). Changes in snow seasonality affect streamflow at the regional scale, with an earlier peak in spring and a possible decrease of low-level flow in summer (Berghuijs et al., 2014; Kang et al., 2016; Dudley et al., 2017), while glacier shrinking can also alter the low-level flow in mountain catchments (Lutz et al., 2014; Milner et al., 2017; Huss and Hock, 2018). This can be partly ameliorated by water management in regulated catchments (Arheimer et al., 2017), but not in large river basins such as the Amazon which also shows an increased seasonality of discharge since 1979 (Liang et al., 2020).

Increasing aridity contrasts between wet and dry seasons over the late 20th century have been suggested (Kumar et al., 2015), with a human-induced decrease of water availability during the dry season over Europe, western North America, northern Asia, southern South America, Australia and eastern Africa (Padrón et al., 2020). Seasonal contrasts in microwave surface soil moisture measurements have also increased over 1979-2016 (Pan et al., 2019). Terrestrial water storage variations derived from gravimetric measurements since 2003 show a strong seasonality which is underestimated by global hydrological models (Scanlon et al., 2019) and whose multidecadal trends are difficult to interpret given the direct effect of enhanced water use (Rodell et al., 2018; Scanlon et al., 2018).

In summary, there is *medium confidence* that the annual range of precipitation has increased since the 1980s, at least in subtropical regions and over the Amazon. There is *low confidence* that this increase is due to human influence and that GHG forcing has already altered the timing or duration of wet seasons. There is *high confidence* that the human-induced retreat of the springtime snow cover and melting of glaciers have already contributed to changes in streamflow seasonality in high-latitude and low-elevation mountain catchments, and *medium confidence* that human activities have also contributed to an increased seasonality of water availability, including a drier dry season, in the extratropics.

Projected changes

AR5 reported with *high confidence* that the contrast between wet and dry seasons will generally increase with global warming and that monsoon onset dates will *likely* become earlier or show little change, while monsoon retreat dates will *likely* be delayed, resulting in a lengthening of the wet season in many regions.

Since AR5, several studies have further documented a projected increase in rainfall seasonality and the understanding of the underlying mechanisms has been improved (Sections 8.2.1 and 8.3.2). CMIP5 models show that the seasonal concentration of annual precipitation will increase over many regions by the end of the 21st century, with robust model agreement in most subtropical regions where an increase in the mean number of dry days was also reported in the RCP8.5 scenario (Pascale et al., 2016). The semi-arid, winter rainfall dominated subtropical climate is projected to shift poleward and eastward, with the equatorward margins replaced by a more arid climate type. However, evolving SST patterns and land-ocean warming contrasts cause more complex responses (Alessandri et al., 2015; Polade et al., 2017; Brogli et al., 2019; Zappa et al., 2020). Projections over California show a stronger and shorter wet season (Polade et al., 2017; Dong et al., 2019). Decreases in future winter and spring rainfall are projected over southwestern Australia (Hope et al., 2015). Central Asia is projected to experience wetter winters, associated with an increase in snow depth in the northeastern regions (Li et al., 2019b). Even in a +2°C climate, both extreme precipitation and dryness will increase significantly in the extratropics, amplifying the seasonal precipitation range (Fujita et al., 2018). A single-model study shows that the annual range of precipitation increases globally by 2.6% per °C of global warming in stabilized low-warming scenarios (Chen et al., 2020f).

In the tropics, an amplified annual cycle (by ~3–5%/°C) of global land monsoon hydroclimates (P, $P - E$, and runoff) is projected by CMIP5 models under the RCP8.5 scenario, mostly due to a more intense wet season (Zhang et al., 2019c). A longer rainy season is projected by CMIP6 models over most regional monsoon areas except in the Americas (Moon and Ha, 2020). A delayed onset and cessation of the wet season over West Africa and the Sahel (Dunning et al., 2018) and a slightly delayed onset of South Asian monsoon rainfall (Hasson et al., 2016) are projected by CMIP5 models. CMIP5 projections suggest a strengthening of the annual cycle and a lengthening of the dry season in Southern Amazonia (Fu et al., 2013; Reboita et al., 2014; Boisier et al., 2015; Pascale et al., 2016; Sena and Magnusdottir, 2020). This is further verified by the projections from 6 CMIP6 models (Moon and Ha, 2020). A wet season shorter by 5-10 days by the end to the 21st century is projected for southern Africa (Dunning et al., 2018).

An increase in streamflow seasonality is projected over several large rivers in the low-mitigation RCP8.5 scenario, but with only small changes in the seasonality timing, except in northern high latitudes due to the earlier but potentially slower snowmelt in a warmer world (Eisner et al., 2017; Musselman et al., 2017). At the end of the century in a high-emission scenario, peak snowmelt timing is projected to occur one month earlier and peak water volume is 79% lower in the eastern USA (Rhoades et al., 2018). Earlier snow melt is projected e.g. by 30 days at the end of the 21st century in RCP4.5 for the Sierra Nevada in the western USA (Sun et al., 2018b). Sub-seasonal changes in water availability were found in many regions in the RCP8.5 scenario. However these should be considered with caution given the magnitude of model errors (Ferguson et al., 2018a). Increases in the seasonality of water availability has been found to be more pronounced in areas with high atmospheric evaporative demand, giving rise to a pattern of seasonally variable regimes becoming even more variable (Konapala et al., 2020). RCP4.5 and RCP8.5 projections show a pronounced soil drying in summer and autumn over western Europe, and a springtime drying over northern Europe due to an earlier snowmelt (Ruosteenoja et al., 2018).

A simple relative seasonality metric (Walsh and Lawler, 1981) applied to global projections based on CMIP6 models and SSP scenarios supports previous CMIP5 findings, especially the amplified seasonality of precipitation around the Mediterranean, and across southern Africa, California, southern Australia and the Amazon (Box 8.2, Figure 1). While such changes are not significant in the low-emission SSP1-2.6 scenario, they are consistent with the increased frequency of dry days projected over the same regions (Figure 8.16). In monsoon regions outside the Americas, rainfall seasonality does not show a significant increase even in high-emission scenarios. This challenges previous CMIP5 findings based on the difference between maximum and minimum monthly precipitation in a year (Zhang et al., 2019c) and higher sensitivity to the projected increase in precipitation extremes (Section 11.4.5). In the northern high-latitudes, milder winters are associated with wetter conditions and a decrease in precipitation seasonality.

In summary, the annual range of precipitation, water availability and streamflow will increase with global warming over subtropical regions and the Amazon (*medium confidence*), especially around the Mediterranean and across southern Africa (*high confidence*). The contrast between the wettest and driest month of the year is *likely* to increase by 3 to 5%/°C with global warming in most monsoon regions, in terms of precipitation, water availability (P-E) and runoff (*medium confidence*). There is *medium confidence* that the monsoon season could be delayed in a warmer climate in the Sahel. There is *high confidence* of earlier snowmelt.

[START BOX 8.2, FIGURE 1 HERE]

Box 8.2, Figure 1: Projected long-term changes in precipitation seasonality. Global maps of projected changes in precipitation seasonality (simply defined as the sum of the absolute deviations of mean monthly rainfalls from the overall monthly mean, divided by the mean annual rainfall as in Walsh and Lawler, 1981) averaged across 31 to 33 CMIP6 models in the SSP1-2.6 (b), SSP2-4.5 (c) and SSP5-8.5 (d) scenario respectively. The simulated 1995-2014 climatology is shown in panel (a). All changes are estimated in 2081-2100 relative to 1995-2014. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END BOX 8.2, FIGURE 1 HERE]

[END BOX 8.2 HERE]

[START FIGURE 8.15 HERE]

Figure 8.15: Projected long-term relative changes in daily precipitation statistics. Global maps of projected seasonal mean relative changes (%) in the number of dry days (i.e. days with less than 1 mm of rain) and daily precipitation intensity (in mm/day, estimated as the mean daily precipitation amount at wet days - i.e., days with intensity above 1 mm/day) averaged across CMIP6 models in the SSP1-2.6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario respectively. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.15 HERE]

[START FIGURE 8.16 HERE]

Figure 8.16: Rate of change in mean and variability across increasing global warming levels. Relative change (%) in seasonal mean total precipitable water (green dashed line), precipitation (red dashed lines), runoff (blue dashed lines), as well as in standard deviation of precipitation (red solid lines) and runoff (blue solid lines) averaged over extra-tropical land in (a) summer and (b) winter, and tropical land in (c) JJA and (d) DJF as a function of global-mean surface temperature for the CMIP6 multi-model mean across the SSP5-8.5 scenario. Extra-tropical winter refers to DJF for Northern Hemisphere and JJA for Southern Hemisphere (and the reverse for extra-tropical summer). Each marker indicates a 21-year period centered on consecutive decades between 2015 and 2085 relative to the 1995–2014 base period. Precipitation and runoff variability are estimated by their standard deviation after removing linear trends from each time series. Error bars show the 5-95% confidence interval for the warmest 5°C global warming level. Figure adapted from (Pendergrass et al., 2017) and updated with CMIP6 models. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.16 HERE]

8.4.1.4 Evapotranspiration

Since AR5, there is a growing body of evidence suggesting that future projections in evapotranspiration are driven by changes in temperature and relative humidity (Laîné et al., 2014; Pan et al., 2015; Ukkola et al., 2016a), as well as precipitation patterns, as found in AR5.

Analysis of CMIP5 models suggests that atmospheric evaporative demand will increase over most areas of the world in high-emission scenarios (*virtually certain*), mostly as a consequence of an increase in vapour pressure deficit (Scheff and Frierson, 2014, 2015; Vicente-Serrano et al., 2020b). CMIP5 models also project an increase in evapotranspiration over most land areas (*medium confidence*) (Laîné et al., 2014). However, regional changes in evapotranspiration can also be influenced by changes in soil moisture and vegetation, which modulate the moisture flux from the land to the atmosphere. Several studies of CMIP5 projections suggest that increases in plant water use efficiency will limit or counteract rising evapotranspiration (Milly and Dunne, 2016; Swann et al., 2016; Lemordant et al., 2018; Yang et al., 2018d). However, other studies have found that transpiration increases due to the impact of climate change on growing season length, leaf area, and evaporative demand (Frank et al., 2015; Mankin et al., 2017, 2018, 2019; Guerrieri et al., 2019; Zhou et al., 2019a; Vicente-Serrano et al., 2020b) (Section 8.2.3.3). The parameterizations accounting for these complex physiological processes in global climate models may also be insufficient (Franks et al., 2017; Peters et al., 2018; Peano et al., 2019). Thus, there is currently *low confidence* in the role of vegetation physiology in modulating future projections of evapotranspiration.

CMIP6 models project a geographical pattern of changes in evapotranspiration similar to previous generation models (Figure 8.17), although the magnitude is generally larger than found for CMIP5 projections (Liu et al., 2020b). There is however a strong seasonality in many regions, with a larger relative increase in the winter season of the Northern Hemisphere and smaller relative changes in the summer (Figure 8.17). Evapotranspiration increases in most land regions, except in areas that are projected to become moisture-limited (due to reduced precipitation and increased evaporative demand), such as the Mediterranean, South Africa, and the Amazon basin (*medium confidence*). The patterns of change increase in magnitude from low to high-emission SSP scenarios (*medium confidence*).

In summary, future projections indicate that anthropogenic forcings will drive an increase in global mean evaporation over most oceanic areas (*high confidence*) (Figure 8.17), an increase in global atmospheric demand (*virtually certain*) and an increase in evapotranspiration over most land areas, with the exception of moisture-limited regions (*medium confidence*). However, substantial uncertainties in projections of evapotranspiration, especially at seasonal and regional scales, remain (see also Section 8.2.3.3, Cross-Chapter Box 5.1).

[START FIGURE 8.17 HERE]

Figure 8.17: Projected long-term relative changes in seasonal mean evapotranspiration. Global maps of projected relative changes (%) in seasonal mean of surface evapotranspiration for DJF (left panels) and JJA (right panels) averaged across 29 or 30 CMIP6 models for SSP1.2-6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario respectively. All changes are estimated in 2081-2100 relative to 1995-2014. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.17 HERE]

8.4.1.5 Runoff, streamflow and flooding

AR5 assessed that projected changes in runoff had *low confidence* over the period 2016-2035; however, under the RCP8.5 scenario, runoff increases by 2100 are *likely* in high northern latitudes. This is consistent with projected regional precipitation increases, and that runoff decreases that are *likely* in southern Europe, the Middle East and southern Africa, based on consistency of changes across different generations of models and different forcing scenarios. There was considerable uncertainty in the magnitude and direction of change for some regions, largely driven by the uncertainty in projected precipitation changes, particularly across south Asia. For flooding, AR5 assessed with *medium confidence* that flooding would increase over parts of South and Southeast Asia, tropical Africa, northeast Eurasia, and South America, and decrease for parts of northern and Eastern Europe, Anatolia, central Asia, central North America, and southern South America. SR1.5 assessed with *medium confidence* that warming of 2˚C would increase the fraction of global area affected by flood hazard relative to warming of 1.5˚C. Projected climate-driven changes to runoff, streamflow, and flooding will occur in the context of potential human-caused land-use and land-cover changes, which can have a large influence on surface water (Sterling et al., 2013) but which have considerable uncertainty in projections (Prestele et al., 2016).

Since AR5, studies confirm that global mean annual runoff increases with global surface temperature increase (Zhang and Tang, 2014; Zhang et al., 2018e; Lehner et al., 2019), but varies regionally (Chen et al., 2017; Yang et al., 2017; Cook et al., 2020). CMIP5 models display a large spread in the ratio of runoff to precipitation for the present-day climate, which applies also to future runoff changes under global warming (Lehner et al., 2019). In studies of CMIP6 projections, runoff increases in most parts of the northern high latitudes and Asia and north and eastern Africa, and decreases in the Mediterranean region, southern Africa, southern Australia and in parts of western Africa, as well as in Central and South America (Greve et al., 2018; Cook et al., 2020). Projected changes in runoff also vary seasonally. In the Northern Hemisphere, runoff increases during winter since more precipitation falls as rain than snow and decreases in the summer as less snow is available to contribute to runoff during the warm season (Cook et al., 2020). Global maps of projected changes for DJF and JJA are shown in in Figure 8.18, showing projected changes becoming larger and more consistent in the higher emissions scenarios. Runoff projections for CMIP6 are also shown in Figure 8.16 for tropical and extratropical averages at a range of global mean warming levels and in Table 8.1 for global land in different future scenarios. In the tropics, both the mean and interannual variability of runoff increase with warming. The increase in variability is roughly twice as large as the increase in the mean, and has a large spread across models. In the extratropics, changes are small in the summer but there are large increases in the winter, with the mean increasing much more than the variability, in contrast to the tropics.

Changes in streamflow vary regionally and increase in magnitude with emissions scenario, as with runoff (although the two are not equivalent, as runoff includes both surface runoff and streamflow). Streamflow projections additionally require the use of hydrologic models forced by the output from climate models and have not been as widely explored as they are not variables directly included in climate models. On an annual basis, streamflows have been projected to increase in the Northern high latitudes and tropical Asia and Africa, and to decrease in the Mediterranean, tropical South America, and South Africa (Döll et al., 2018b). For a 4°C global warming, half of the global land area is projected to be exposed to increased high flows (average increase 25%), while about 60% may be exposed to decreased low flows (average decrease 50%) (Asadieh and Krakauer, 2017).

Changes in the seasonality of runoff and streamflow are assessed in Box 8.2. The seasonality of runoff and streamflow (calculated as the annual difference between the wettest and driest months of the year), is expected to increase with global warming in the subtropics, especially in the Mediterranean and southern Africa with *high confidence*, and in the Amazon with *medium confidence*. For regions where snowmelt is an important contributor to streamflow, there is *high confidence* that snowmelt occurring earlier in the year will result in peak flows also occurring earlier in the year, and *medium confidence* that reduced snow volume and the weaker solar radiation earlier in the year will reduce the most intense flows. In roughly half of 56 large-scale glacierized drainage basins, projected runoff changes show an increase until a maximum is reached, beyond which runoff steadily declines because of limited ice volumes (Huss and Hock, 2018).

As future changes in flood events are assessed in Chapters 9 and 11, only a summary is presented here. There are a number of complicating factors for projecting both pluvial (overland) and fluvial (river) flooding that limit confidence in their assessment. In addition to precipitation, flooding also depends on basin and river characteristics such as permeability, antecedent soil moisture, and antecedent flow levels for river flooding, so projections of extreme precipitation and flooding are not always closely linked (Section 8.2.3.2).

Possible changes in water resources management and land use add another layer of complexity to future changes. There is *medium confidence* in a general increase in pluvial and fluvial flooding, although there are large geographical variations in magnitude. There are increases in flooding in the West Amazon, the Andes, and northern Eurasia (Chapter 11, Section 11.5.5). There is *medium confidence* in future increases in urban and coastal floods (Chapter 11, Section 11.5.5), and *high confidence* that some coastal regions will experience large increases in surge flooding (Chapter 9, Section 9.6.4.2). There is *medium confidence* in an increase in compound flood events (Chapter 11, Section 11.8.1). Although there is currently insufficient evidence for a confident projection, flooding due to rain-on-snow events can be expected to decrease where snow decreases (Chapter 11, Section 11.8.3), and the seasonality of snowmelt-related flooding can be expected to shift in regions with temperature-driven shifts in the snowmelt season (e.g., Vormoor et al., 2015). Glacier lake outburst floods (GLOFs) are expected to increase substantially, in delayed response to glacier recession but with *low confidence*, due to the small number of studies and the complexity of the processes involved (Chapter 9, Section 9.5.3.3).

In summary, there is *medium confidence* that global runoff will increase with global warming, but with large regional and seasonal variations. There is *high confidence* that runoff will increase in the northern high latitudes and decrease in the Mediterranean region and southern Africa. There is *medium confidence* that runoff will increase in regions of central and eastern Africa, and decrease in Central America and parts of southern South America, with the magnitude of the change increasing with emissions. There is *medium confidence* that the seasonality of runoff and streamflow will increase with global warming in the subtropics. In snow-dominated regions, there is *high confidence* that peak flows associated with spring snowmelt will occur earlier in the year and *medium confidence* that snowmelt-induced runoff will decrease with reduced snow, except in glacier-fed basins where runoff may increase in the near term. There is *medium confidence* that flooding in general will increase, although with considerable variation based on geographic region and flood type. These projected climate-related changes will occur in the context of human-caused land-use and land-cover changes, which may also have a large influence.

[START FIGURE 8.18 HERE]

Figure 8.18: Projected long-term relative changes in seasonal mean runoff. Global maps of projected relative change (%) in runoff seasonal mean for DJF (left panels) and JJA (right panels) averaged across CMIP6 models SSP1.2-6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario respectively. All changes are estimated in 2081-2100 relative to 1995-2014. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.18 HERE]

8.4.1.6 Aridity and drought

AR5 concluded that regional to global-scale projections of aridity and drought remained relatively uncertain compared to other aspects of the water cycle. It reported that there is a *likely* increase in drought occurrence (*medium confidence*) by 2100 in regions that are currently drought-prone under the RCP8.5 scenario due to projected decreases in soil moisture. It stated that it is *likely* that the most prominent projected decreases in soil moisture would occur in the Mediterranean, southwest USA, and southern Africa, consistent with projected changes in the Hadley Circulation and increased surface temperatures. These AR5 conclusions are generally supported by more recent analyses of CMIP5 models (Feng and Fu, 2013; Berg et al., 2017; Cook et al., 2018).

Results from the latest generation of models in CMIP6 are largely congruent with CMIP5. Consistent with the coherent nature of warming in future projections, increases in vapour pressure deficit and evaporative demand are widespread and consistent across regions, seasons, and models, increasing in magnitude in accordance with the emissions scenario (Figure 8.19) (*high confidence*) (Scheff and Frierson, 2014, 2015; Vicente-Serrano et al., 2020b). Even under a low-emissions scenario (SSP1-2.6), projections of soil moisture show significant decreases in the Mediterranean, southern Africa, and the Amazon basin (*high confidence*) (Figure 8.19). Under mid- and high-emissions scenarios (SSP2-4.5 and SSP5-8.5), coherent declines emerge across Europe, westernmost North Africa, southwestern Australia, Central America, southwestern North America, and southwestern South America (*high confidence*) (Figure 8.19) (Cook et al., 2020). Compared to CMIP5 results, CMIP6 models exhibit more consistent drying in the Amazon basin (Parsons, 2020), more extensive declines in total soil moisture in Siberia (Cook et al., 2020), and stronger declines in westernmost North Africa and southwestern Australia (Figure 8.19).

Soil moisture in the top soil layer (10 cm) shows more widespread drying than total soil moisture, reflecting a greater sensitivity of the upper soil layer to increasing evaporative demand (Berg et al., 2017) (Figure 8.19). Conversely, total column soil moisture represents the carry-over of moisture from previous seasons deeper in the soil column, and potentially higher sensitivity to vegetation processes (Berg et al., 2017; Kumar et al., 2019). Central America, the Amazon basin, the Mediterranean region, southern Africa, and southwestern Australia are projected to experience significant declines in total soil moisture, whereas declines in Europe (north of the Mediterranean), western Siberia, and northeastern North America are limited to the surface (Figure 8.19). It should be noted that because models differ in their number of hydrologically active layers, there is less confidence in total soil moisture projections than surface soil moisture projections. Based on surface soil moisture projections, more than 40% of global land areas (excluding Antarctica and Greenland) are expected to experience robust year-round drying, even under lower emissions scenarios (Cook et al., 2020). The percentage of land area experiencing drying is slightly lower when runoff is used as an aridity metric instead (20–30%); taking this into consideration, it is estimated that about a third of global land areas will experience at least moderate drying in response to anthropogenic emissions, even under SSP1-2.6 (*medium confidence*) (Cook et al., 2020).

Although there are regions where multiple models predict consistent and significant changes in soil moisture, as with evapotranspiration (Section 8.4.1.4), there is still uncertainty in these projections related to the response of plants to elevated CO_2. Most models project increases in two variables that have opposite effects on surface water availability: plant water use efficiency (WUE) and leaf area index (LAI) (see Section 8.4.1.4). As discussed in Section 8.2.3.3, 8.3.1.4, and 8.4.1.4, there is *low confidence* in how these changes in plant physiology will affect future projections of evapotranspiration, and likewise, drought and aridity.

[START FIGURE 8.19 HERE]

Figure 8.19: Projected long-term relative changes in annual mean soil moisture and vapour pressure deficit. Global maps of projected relative changes (%) in annual mean vapour pressure deficit (left), surface soil moisture (top 10cm, middle) and total column soil moisture (right) from available CMIP6 models for the SSP1.2-6 (a,b,c), SSP2-4.5 (d,e,f) and SSP5-8.5 (g,h,i) scenarios respectively. All changes are estimated for 2081-2100 relative to a 1995-2014 base period. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement ("Robust change"), where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.19 HERE]

Changes in meteorological (precipitation-based) drought duration and intensity in CMIP6 models are more robust than projected changes in mean precipitation, more than found in CMIP5 projections (Ukkola et al., 2020). Significant increases in drought duration are expected in Central America, the Amazon basin, southwestern South America, the Mediterranean, westernmost North Africa, southern Africa, and southwestern Australia, on the order of 0.5–1 month for a moderate emissions scenario (SSP2-4.5) and 2 months for a high emissions scenario (SSP5-8.5) (Ukkola et al., 2020). Drought intensity is projected to increase in the tropics, mainly in the Amazon basin, central Africa, and southern Asia, as well as in Central America and southwestern South America (Ukkola et al., 2020). The CORDEX South Asia multi-model ensemble projections indicate an increase in the frequency and severity of droughts over central and northern India during the 21st century, under the RCP4.5 and RCP8.5 scenarios *(medium confidence)* (Mujumdar et al., 2020). Under middle or high emissions scenarios, the likelihood of extreme droughts (events that have magnitudes equal to or less than the 10th percentile of the 1851–1880 baseline period) increases by 200–300% in the Amazon basin, southwestern North America, Central America, the Mediterranean, southern Africa, and southwestern South America (Cook et al., 2020). Even under a low emissions scenario (SSP1-2.6), the likelihood of extreme droughts increases by 100% in southwestern North America, southwestern South America, the Amazon, the Mediterranean, and southern Africa (Cook et al., 2020). Thus, there is *high confidence* that drought severity and intensity will increase in the Mediterranean, southern Africa, southwestern South America, southwestern North America, southwestern Australia, Central America and the Amazon basin.

Paleoclimate records provide context for these future expected changes in drought and aridity. In the Mediterranean, western North America, and central Chile, there is *high confidence* that climate change will shift soil moisture (as represented by the Palmer Drought Severity Index) outside the range of observed and reconstructed values spanning the last millennium (Cook et al., 2014; Otto-Bliesner et al., 2016) (Figure 8.20). Warmer temperatures, leading to increased evaporative losses, are clearly implicated in the projected future drying in these semi-arid regions (Dai et al., 2018), emphasizing the central role that warming plays in driving increased evaporative demand (Vicente-Serrano et al., 2020b). In contrast, future trajectories are more uncertain in regions like central Asia and eastern Australia–New Zealand where projected changes in precipitation and soil moisture are less coherent (Hessl et al., 2018) (Figure 8.19, 8.20). More information on projected changes in drought, including specific categories or drought, can be found in Section 11.6.5 and Section 12.4.

In summary, there is *high confidence* that soil moisture will decline in semi-arid, winter-rainfall dominated areas including the Mediterranean, southern Africa, southwestern North America, southwestern South America, and southwestern Australia, as well as in Central America and the Amazon basin. In general, these regions are expected to become drier both due to reduced precipitation (*medium confidence*) and increases in evaporative demand (*high confidence*). These same regions are *likely* to experience increases in drought duration and/or severity (*high confidence*). The magnitude of expected change scales with emissions scenarios (*high confidence*) but even under low-emissions trajectories, large changes in drought and aridity are expected to occur (*high confidence*) with consequences for regional water availability. In the Mediterranean, central Chile, and western North America, future aridification will far exceed the magnitude of change seen over the last millennium (*high confidence*).

[START FIGURE 8.20 HERE]

Figure 8.20: Past-to-future drought variability in paleoclimate reconstructions and models for select regions. On the left (a,c,e,g,i), tree-ring reconstructed Palmer Drought Severity Index (PDSI) series (black line) for the Mediterranean (10˚W–45˚E, 30˚–47˚N; Cook et al., 2015, 2016), central Chile (70˚–74˚W, 32˚–37˚S; Morales et al., 2020), western North America (117˚–124˚W, 32˚–38˚N; Cook et al., 2010; Griffin and Anchukaitis, 2014), Eastern Australia and New Zealand (136˚–178˚E, 46˚–11˚S; Palmer et al., 2015), and Central Asia (99˚–107˚E, 47˚–49˚N; Pederson et al., 2014; Hessl et al., 2018) plotted in comparison to the past-to-future fully-forced simulations from four ensemble members (thin blue lines) from the NCAR CESM Last Millennium Ensemble (thick blue line = ensemble mean) (Otto-Bliesner et al., 2016) for the same regions. The shaded area represents the range (10th to 90th percentile) of historical and future (RCP8.5) PDSI (Penman-Monteith) simulations from 15 CMIP5 models and 34 ensemble members for the same regions (1900–2100; Cook et al., 2014). On the right (b,d,f,h,i), the distribution of annual PDSI values from the past and present (850 to 2005 CE) (black) is compared to the future distribution (2006 to 2100 CE) (blue). The distributions show each of the four ensemble members from the CESM LME simulations. The future component of the CESM LME follows the RCP8.5 scenario. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.20 HERE]

8.4.1.7 Freshwater reservoirs

8.4.1.7.1 Glaciers

Previous assessments have concluded that recent warming has led to a reduction in low-elevation snow cover (SROCC, *high confidence*), permafrost (SROCC, *high confidence*), and glacier mass (AR5, *high confidence*; SROCC, *very high confidence*). SROCC noted that these declines are projected to continue almost everywhere over the 21st century (*high confidence*), with complete glacier loss expected in regions with only small glaciers (*very high confidence*). SROCC supported the AR5 finding that glacier recession would continue even without further changes in climate. SROCC concluded that cryosphere changes had already altered the seasonal timing and volume of runoff (*very high confidence*), which in turn had affected water resources and agriculture (*medium confidence*), and projected peak water runoff had already been reached before 2019 in some of the glacier regions considered.

Chapter 9 provides detailed assessment of glacier observations and projections (Figure 9.20 and 9.21, see Section 9.5.1). Here, a summary of their key findings is presented. Since SROCC, the coordinated glacier model intercomparison project (GlacierMIP (Marzeion et al., 2020), see also Box 9.3) has advanced modelling efforts. Global glacier volumes will substantially decline in coming decades regardless of emissions scenario; under a high emission scenario some areas will lose nearly all of their glacier mass (Section 9.5.1.3). The projected global glacier mass loss over 2015-2100 is 29 000 ± 20 000 Gt for SSP1-2.6 to 58 000 ± 30 000 Gt for SSP5-8.5 (Section 9.5.1). Because of their lagged response to warming, glaciers will continue to lose mass for decades even if global temperature is stabilized (*very high confidence*) (Section 9.5.1).

Global glacier mass loss projections show a scenario-dependent geographic partitioning of when peak in runoff occurs (Marzeion et al., 2020), consistent with previous studies (Radić et al., 2014; Huss and Hock, 2018; Hock et al., 2019a). Under a low emission scenario (Marzeion et al., 2020) all regions exhibit runoff in the decades prior to 2050. Under a high emission scenario however, low and mid-latitude regions show peak runoff before approximately 2060, whereas Arctic regions peak in later decades around 2070-2090. Antarctic glacier losses will not have peaked by the end of the century in the high emission scenario. Globally, peak runoff of 2.5 to 3 mm / year sea level equivalent occurs around 2090 (Marzeion et al., 2020). Regional projections are presented in detail in Section 9.5.1 and Figure 9.21, and briefly summarized below.

Himalaya and Central Asia: Glaciers in the Himalayas feed ten of the world's most important river systems and are critical water sources for nearly two billion people (Wester et al., 2019) However they are some of the most vulnerable 'water towers' (Immerzeel et al., 2020) that are projected to experience volume losses of approximately 30 to 100% by 2100 depending on global emissions scenarios (Marzeion et al., 2020). Under mid-range emissions scenarios glaciers in this region are projected to reach peak runoff during the period 2020 to 2040 (Marzeion et al., 2020).

Alaska, Yukon, British Columbia: Post-AR5 but pre-SROCC projections indicated a potential 70 ± 10% reduced volume of glacier ice in western Canada relative to 2005 (Clark et al., 2015), with few glaciers remaining in the Interior and Rockies regions and maritime glaciers in northwestern British Columbia surviving only in a diminished state. Recent global projections support these earlier findings, showing that glacier mass in Western Canada and the United States may reduce by 50% under low emissions scenarios and be completely lost under the highest emissions and most sensitive glacier model combinations (Marzeion et al., 2020) (Figure 9.21). Arctic Canada and Alaskan glaciers are projected to experience more modest mass loss (0 to 60% depending on region, scenario, and model; Marzeion et al., 2020).

Andes: Huss and Hock (2018) concluded that peak glacier mass was reached prior to 2019 for 82–95% of the glacier area in the tropical Andes. This is consistent with more recent global model simulations that show mass loss rates from low latitude glaciers that universally decline from the start of simulations in 2015, regardless of emissions scenario (Marzeion et al., 2020). Peak runoff in low-latitude Andean glacier-fed rivers has therefore already passed (Frans et al., 2015; Polk et al., 2017) but in the Southern Andes may occur in the latter half of the century under high emission scenarios (Marzeion et al., 2020).

In summary, glaciers are projected to continue to lose mass under all emissions scenarios *(very high confidence*). Runoff from glaciers is projected to peak at different times in different places, with maximum rates of glacier mass loss in low latitude regions taking place in the next few decades in all scenarios (*high confidence*). While runoff from small glaciers will typically decrease because of glacier mass depletion, runoff from larger glaciers will increase with increasing global warming until glacier mass is similarly depleted, after which runoff peaks and then declines and which tends to occurs later in basins with larger glaciers and higher ice-cover fractions (*high confidence*). Glaciers in the Arctic and Antarctic will continue to lose mass through the latter half of the century and beyond (*high confidence*).

8.4.1.7.2 Seasonal snow cover

AR5 assessed as *very likely* that the amount and seasonal duration of Northern Hemisphere snow cover will reduce under global warming (AR5 Section 11.3.4.2, Section 12.4.6.2). Changes in the total amount of water in the snow cover (snow water equivalent) are less certain because of the competing influences of temperature and precipitation.

As snow cover is assessed in Chapter 9 (Section 9.5.3.3), only an overview of that assessment is provided here. Changes in seasonality of snow cover are assessed in Box 8.2. The continued consistency of reported results across all generations of model projections, along with improvements in process understanding, has increased confidence in snow cover projections since AR5.

In summary, based on the results of Chapter 9, it is now *virtually certain* that future Northern Hemisphere snow cover extent and duration will continue to decrease with global warming. While most studies have focused on the Northern Hemisphere, process understanding suggests with *high confidence* that these results apply to the Southern Hemisphere as well. There is *high confidence* in snowmelt occurring earlier in the year. Changes to the timing and amount of snowmelt will have a strong influence on all the other aspects of the water cycle in regions with seasonal snow, including run-off, soil moisture, and evapotranspiration.

8.4.1.7.3 Wetlands and lakes

AR5 did not include specific projections for wetlands and lakes. SRCCL and SROCC provided some discussion of wetlands projections. For coastal wetlands, SRCCL noted the importance of sea level rise for increased saltwater intrusion, although projections of coastal wetland area with sea level rise are inconclusive. Some studies project substantial decreases (Spencer et al., 2016) while others indicate possible increases (Schuerch et al., 2018). SRCCL also noted the general expectation for decreases in water resources, including wetlands, in areas of decreased rainfall due to increased evaporation.

Local studies of inland wetlands project decreases in a range of environments including mountain (Lee et al., 2015), mid-to-high latitude (Zhao et al., 2018b), and prairie (Sofaer et al., 2016) regions. In addition to affecting wetland extent and density, changes in flooding can also affect the connectivity between wetlands and rivers (Karim et al., 2016). Despite a number of uncertainties underlying the general response of wetlands to climate change, there are multiple ways climate change may cause considerable stress on both inland and coastal wetlands (Junk et al., 2013; Moomaw et al., 2018).

Widespread changes are also projected for lakes (Woolway et al., 2020), including changes in lake temperature (Fang and Stefan, 1999; Sahoo et al., 2016), ice (Sharma et al., 2019), evaporation (Wang et al., 2018d), and stability and mixing (Woolway and Merchant, 2019). Note that lake ice is also considered in Chapter 12 of this Report. To date, carbon dioxide-induced lake acidification, analogous to ocean acidification, has not been the focus of many studies but may occur with continued emissions (Phillips et al., 2015). While glacier lakes in general increase with melting glaciers (Linsbauer et al., 2016; Colonia et al., 2017; Magnin et al., 2020) no clear projections are currently available (see discussion in Chapter 9). Projections of lake level means and variability show substantial changes for individual lakes (Bucak et al., 2017; Li et al., 2021) but can be sensitive to methodology, due to the competing processes involved (Notaro et al., 2015). Projected changes to wetlands and lakes due to climate change will occur in the context of widespread and continuing human-caused conversion and degradation of wetlands (e.g, Davidson, 2014), and where water withdrawals have a large impact on lake levels (e.g., (Micklin, 2016)).

In summary, there is *medium confidence* that inland wetland extent will decrease in regions of projected precipitation decrease and evaporation increase, and *high confidence* that sea level rise will increase saltwater intrusion into coastal wetlands. However, there is *low agreement* on the influence of sea level rise on the extent of coastal wetlands. Regarding lakes, there is *high confidence* for temperature increases and ice decreases, based on both projections and physical expectations, and *low confidence* for non-homogeneous decreases in mixing, given there is currently *limited evidence*.

8.4.1.7.4 Groundwater

Groundwater projections were not assessed in AR5. Groundwater processes are not explicitly included in most current CMIP6 models and so must be calculated separately with hydrologic models (e.g., Taylor et al., 2013; Cuthbert et al., 2019a). A range of factors are important in assessing groundwater projections, including the mean difference between precipitation and evaporation, the intensity of precipitation (Taylor et al., 2013b), and in changes in snow (Tague and Grant, 2009), glaciers (Gremaud et al., 2009), and permafrost (Okkonen and Kløve, 2011). Climate impacts on groundwater are occurring in the context of severe and growing human-caused groundwater depletion (Konikow and Kendy, 2005), Rodell et al. 2018; (Bierkens and Wada, 2019), also see WGII), and water scarcity issues (Mekonnen and Hoekstra, 2016). Climate-related changes to the water cycle can influence water demand (for example, precipitation decreases in an irrigated area), and anthropogenic groundwater depletion can influence the water cycle through interactions with surface energy fluxes, surface water, and vegetation (Cuthbert et al., 2019a), although uncertainties in estimates of future groundwater depletion are large ((Smerdon, 2017), (Bierkens and Wada, 2019)). Some aspects of groundwater change will be irreversible, including the increase of saltwater intrusion into coastal aquifers with sea level rise (Werner and Simmons, 2009), and depletion of fossil aquifers and aquifers with very long recharge times ((Bierkens and Wada, 2019)).

Globally, two modelling studies have shown substantial decreases in groundwater in regions including the Mediterranean, northeastern Brazil and southwestern Africa, with less clarity for other regions (Döll, 2009), Portmann et al., 2013). Recent regional-scale analyses of the impact of water cycle changes on groundwater recharge (e.g. Meixner et al., 2016; Shrestha et al., 2018; Tillman et al., 2017) suggest changes in both seasonality and spatial distribution, which are amplified under a higher greenhouse-gas emission scenario (i.e., RCP 8.5 compared to RCP4.5). Seasonality changes are linked to increases during wet winter periods and declines during dry summer periods. Changes in spatial distribution are linked with increases in more humid regions and declines in more arid locations. Uncertainty in projections of groundwater were found to be substantially influenced by the conceptual and numerical models employed to estimate groundwater recharge (Meixner et al., 2016; Hartmann et al., 2017). Accordingly, current research on estimating water cycles change on groundwater includes a focus on improving the numerical representation of groundwater systems (Bierkens et al., 2015; Döll et al., 2016)).

In summary, based on known limitations in current modelling, no confident assessment of groundwater projections is made here, although important climate-related changes in groundwater recharge are expected. In many environments, such climate-related impacts are expected to occur in the context of substantial human groundwater withdrawals depleting groundwater storage.

8.4.2 Projected changes in large scale phenomena and regional variability

A weakening of the tropical circulation represents a balance between thermodynamic increases in low level water vapour (~7%/K) and smaller increases in global precipitation (1-3%/K) that are influenced by rapid adjustments to radiative forcings as well as slow responses to warming (Bony et al., 2013; Chadwick et al., 2013; Ma et al., 2018, Section 8.2.2.2). Since AR5, additional drivers of tropical circulation weakening have been identified, including mean SST warming and changes in spatial patterns of SST (He and Soden, 2015), and the direct CO_2 radiative effect (Bony et al., 2013; Merlis, 2015 ; He and Soden, 2015).

8.4.2.1 ITCZ and tropical rain belts

CMIP5 projections show no consistent shift in the zonal mean position of the ITCZ (Byrne et al., 2018; Donohoe et al., 2013; Donohoe and Voigt, 2017). The ITCZ position is strongly connected to cross-equatorial energy transport (Bischoff and Schneider, 2014; Kang et al., 2008), which also shows no consistent change in future projections (Donohoe et al., 2013). Since AR5 it has been reported that most CMIP5 models project a narrowing of the ITCZ in response to surface warming together with intensified ascent in the core region and weakened ascent on the ITCZ edges (Lau and Kim, 2015; Byrne et al., 2018), implying a narrowing of precipitation regions influenced by the ITCZ. Modelled changes in the width and intensity of the zonal mean ITCZ are strongly anti-correlated, i.e. narrowing is associated with increased intensity while broadening with decreased intensity. Such changes are associated with changes in tropical high cloud fraction and outgoing longwave radiation (Su et al., 2017; Byrne et al., 2018).

Regional shifts in tropical convergence zones are much larger than their zonal mean, and associated regional changes in precipitation (Chadwick et al., 2013; Mamalakis et al., 2021) are characterized by considerable uncertainties across models (Kent et al, 2015; Oueslati et al., 2016). Over the tropical oceans, shifts in rain bands are strongly coupled with changes in SSTs (Xie et al., 2010; Huang et al., 2013). Over tropical land, factors including remote SST increases (Giannini, 2010), the direct CO_2 effect (Biasutti, 2013) and land-atmosphere interactions (Chadwick et al., 2017; Kooperman et al., 2018) influence projections. CMIP6 models project a clear northward ITCZ shift over eastern Africa and the Indian Ocean and a southward shift over the eastern Pacific and Atlantic oceans, as a result of regionally-contrasting inter-hemispheric energy flows (Mamalakis et al., 2021). The northward movement of the ITCZ over Africa has been linked to an intensification of the Saharan heat low associated with greenhouse gas warming (Dong and Sutton, 2015), causing the tropical rain belt to seasonally migrate farther northward and reside there longer (Cook and Vizy, 2012; Dunning et al., 2018). In southern Africa, the projected delay in the wet season onset (Dunning et al., 2018) is also associated with a circulation-based northward shift in the tropical rain band (Lazenby et al., 2018).

In summary, consistent with the AR5, the overall weakening of the tropical circulation is projected in CMIP5 and CMIP6 simulations with *high confidence.* It is *likely* that the zonal mean of the ITCZ will narrow and strengthen in the core region with projected surface warming (*high confidence*). Distinct regional shifts in the ITCZ will be associated with regional changes in precipitation amount and seasonality (*medium confidence*).

8.4.2.2 Hadley Circulation and subtropical belt

AR5 found that the Hadley cells are *likely* to slow down and expand in response to radiative forcing, but with considerable internal variability. Given the complexities in forcing mechanisms, AR5 assigned *low confidence* to near-term changes in the structure of the Hadley circulation. The widening Hadley cells were expected to result in a poleward expansion of subtropical dry zones.

Model simulations since AR5 project a more noticeable and consistent weakening of the Northern Hemisphere winter Hadley cell than the Southern Hemisphere winter cell (Seo et al., 2014; Zhou et al., 2016), related to changes in meridional temperature gradient, static stability, and tropopause height (Seo et al., 2014; D'Agostino et al., 2017). Changes in SST patterns reduces the magnitude of Hadley cell weakening (Gastineau et al., 2009; Ma et al., 2012). There is considerable structure in Hadley circulation strength changes with longitude, associated with cloud-circulation interactions (Su et al., 2014). Subtropical anticyclones are projected to intensify over the north Atlantic and south Pacific but to weaken elsewhere (He et al., 2017).

A consistent poleward expansion of the edges of the Hadley cells is projected (Nguyen et al., 2015; Grise and Davis, 2020), particularly in the Southern Hemisphere, consistent with observed trends (Nguyen et al., 2015) (Fig 8.21, 8.3.2.2). The main driver of future expansion appears to be greenhouse gas forcing (Grise et al., 2019), with uncertainty in magnitude due to internal variability (Kang et al., 2013). Proposed mechanisms for poleward expansion include increased dry static stability (Lu et al., 2007 ; Frierson et al., 2007), increased tropopause height (Chen and Held, 2007; Chen et al., 2008), stratospheric influences (Kidston et al., 2015) and radiative effects of clouds and water vapour (Shaw and Voigt, 2016, see also 4.5.1.5). Hadley cell expansion is thought to be associated with the precipitation declines projected in many subtropical regions (Shaw and Voigt, 2016), but more recent work suggests that these reductions are mainly due to the direct radiative effect of CO_2 forcing (He and Soden, 2015), land-sea contrasts in the response to forcing (Shaw and Voigt, 2016; Brogli et al., 2019) and SST changes (Sniderman et al., 2019). In semi-arid, winter rainfall-dominated regions (such as the Mediterranean), thermodynamic processes associated with the land-sea thermal contrast and lapse rate changes dominate the projected precipitation decline in summer, whereas circulation changes are of greater importance in winter (Brogli et al., 2019). The hydroclimates in these regions are projected to evolve with time due to changing contributions from rapid atmospheric circulation changes and their associated SST responses, as well as slower SST responses to anthropogenic forcing (Zappa et al., 2020).

In summary, CMIP5 and CMIP6 models project a weakening of the Hadley cells, with *high confidence* for the Northern Hemisphere in boreal winter and *low confidence* for the Southern Hemisphere in austral winter. The Hadley cells are projected to expand polewards with global warming, most notably in the Southern Hemisphere (*high confidence*). There is currently *low confidence* in the impacts on regional precipitation in subtropical regions.

8.4.2.3 Walker circulation

AR5 determined that the Pacific Walker circulation was *likely* to slow down over the 21st century, which would lead to decreased precipitation over the western tropical Pacific and increases over the central and eastern Pacific. Recent studies show consistency with AR5 conclusions but also show an eastward shift over the Pacific, mostly due to a shift towards more "El Niño–like" conditions under global warming (Bayr et al., 2014). Other studies suggest that the weakening of the Walker circulation is related to the response of the western North Pacific monsoon and to changing land–sea temperature contrasts, while a positive ocean–atmosphere feedback amplifies the weakening of both east–west SST gradient and trade winds in the tropical Pacific (Zhang & Li, 2017).

Since AR5, the paradox between the projected weakening and the observed strengthening of the Walker circulation since the 1990s (Section 8.3.2.2) has triggered debate about the drivers of these changes (England et al., 2014; McGregor et al., 2014; Kociuba and Power, 2015; Vilasa et al., 2017; Chung et al., 2019). Projected changes in equatorial SST gradients are not entirely consistent with observed trends (Coats and Karnauskas, 2017; Seager et al., 2019a), and one CMIP5 model that projects a future strengthening of the Walker circulation is more consistent with observations than other models (Kohyama et al., 2017). Other studies suggest that these differences arises from the dominant influence of internal climate variability to the observed trends (Chung et al., 2019), or as a consequence of a systematic cold bias of most CMIP5 models in their Equatorial Pacific cold tongues (Seager et al., 2019a). However, the latter hypothesis is based on a simplified model of tropical Pacific dynamics and is not consistent with the current physical understanding of the tropical circulation response to increasing CO_2 levels (Section 8.2.2.2) or with independent paleoclimate evidence suggesting a weaker Walker circulation under warmer climates (Tierney et al., 2019; McClymont et al., 2020). Different time scales of the tropical Pacific responses to global warming have been highlighted by numerical experiments with both comprehensive and simplified models. Results suggest a transient strengthening of the Walker circulation related to Indian Ocean warming (Zhang et al., 2018b), followed by a slower weakening linked to a strengthened eastern Pacific cold tongue warming emerging after 50-100 years (Heede et al., 2020, Section 7.4.4.2.1).

CMIP6 projections provide further evidence of a significant long-term weakening of the Walker circulation (Fig. 8.21). For instance, a pronounced weakening of the upper-level tropical easterly jet is projected both over the Indian Ocean and tropical eastern Pacific, where declines are projected to exceed 70% by 2100 in the high-emission SSP5-8.5 scenario (Huang et al., 2020a). CMIP6 models agree on a future decrease of the equatorial zonal temperature gradient (Fredriksen et al., 2020), which can lead to weaker trade winds over the tropical Pacific. However, CMIP6 models show a diversity of SST warming patterns in the tropical Pacific (Freund et al., 2020), which contributes to uncertainties in the response of both Walker circulation and ENSO to continued warming.

In summary, there is *high confidence* that the Pacific Walker circulation will weaken by the end of the 21st century, and will be associated with decreased precipitation over the western tropical Pacific and increases farther east. Discrepancies between observed and simulated changes in SSTs in the tropics indicate that a temporary strengthening of the Walker Circulation can arise from a transient response to GHG radiative forcing (*low confidence*) and from internal variability (*medium confidence*).

[START FIGURE 8.21 HERE]

Figure 8.21: Schematic depicting large-scale circulation changes and impacts on the regional water cycle. The central figures show precipitation minus evaporation (P-E) changes at 3°C or global warming relative to a 1850-1900 base period (mean of 23 CMIP6 SSP5-8.5 simulations). Annual mean changes (large map) include contours depicting control climate P-E=0 lines with the solid contour enclosing the tropical rain belt region and dashed lines representing the edges of subtropical regions. Confidence levels assess understanding of how large-scale circulation change affect the regional water.

[END FIGURE 8.21 HERE]

8.4.2.4 Monsoons

In AR5, monsoon precipitation over land was projected to intensify by the end of the 21st century, due to thermodynamic increases in moisture convergence despite weakening of the tropical circulation (see Section 8.2.1.3). Following the definition of regional monsoons in Annex V and Figure 8.11, and the assessment of the observed changes (Section 8.3.2.4), here we provide an assessment of projected changes in regional monsoons. Assessment is provided either in terms of SSP and RCP scenarios and global warming levels available since AR5, or from the newly available CMIP6 projections (Figure 8.22 and Table 8.2). Table 8.2 provides projected changes across the five SSPs used in this report for precipitation (mm/day), P-E (mm/day) and runoff (mm/day) over the regional monsoons for the mid (2041-2060) and long term (2081-2100).

Table 8.2: Monsoon mean water cycle projections in the medium term (2041-2060) and long term (2081-2100) relative to present day (1995-2014), showing present day mean and 90% confidence range across CMIP6 models (historical experiment) and projected mean changes and the 90% confidence range across the same set of models and a range of shared socioeconomic scenarios. All statistics are in units of mm/day. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

		Mid term: 2041–2061 minus reference period					**Long term: 2081–2100 minus reference period**				
	1995-2014 reference period	**SSP1-1.9**	**SSP1-2.6**	**SSP2-4.5**	**SSP3-7.0**	**SSP5-8.5**	**SSP1-1.9**	**SSP1-2.6**	**SSP2-4.5**	**SSP3-7.0**	**SSP5-8.5**
South and Southeast Asian Monsoon (JJAS)											
Precipitation	8.42 [6.66-10.14]	0.44 [0.08-0.74]	0.47 [0.1-0.96]	0.42 [0.03-0.81]	0.32 [-0.08-0.94]	0.54 [0.11-1.18]	0.46 [0.16-0.7]	0.52 [0.13-1.09]	0.66 [0.16-1.1]	0.94 [0.3-1.78]	1.46 [0.66-2.49]
Runoff	3.75 [1.8-5.71]	0.23 [0.1-0.38]	0.29 [0.02-0.65]	0.29 [-0.0-0.66]	0.24 [-0.04-0.52]	0.38 [0.07-0.78]	0.19 [-0.02-0.35]	0.29 [-0.04-0.65]	0.42 [0.04-0.83]	0.7 [0.12-1.2]	1.14 [0.36-2.05]
P-E	5.19 [3.68-6.5]	0.28 [0.03-0.52]	0.36 [-0.0-0.76]	0.36 [0.02-0.69]	0.3 [-0.04-0.85]	0.45 [0.06-0.95]	0.27 [0.06-0.38]	0.38 [0.11-0.76]	0.51 [0.02-0.83]	0.81 [0.24-1.56]	1.15 [0.45-1.84]
East Asian Monsoon (JJA)											
Precipitation	5.59 [4.47-6.86]	0.37 [-0.09-0.93]	0.37 [-0.09-0.87]	0.34 [0.05-0.76]	0.22 [-0.16-0.88]	0.43 [0.03-1.1]	0.43 [0.07-1.02]	0.44 [-0.0-1.08]	0.51 [0.11-1.09]	0.59 [0.02-1.31]	0.84 [0.24-1.74]
Runoff	2.24 [1.28-3.41]	0.11 [-0.16-0.4]	0.13 [-0.19-0.42]	0.13 [-0.15-0.4]	0.15 [-0.29-0.76]	0.2 [-0.11-0.72]	0.16 [-0.08-0.49]	0.16 [-0.13-0.58]	0.22 [-0.13-0.64]	0.36 [-0.05-0.87]	0.51 [0.06-1.24]
P-E	2.41 [1.51-3.31]	0.1 [-0.31-0.51]	0.13 [-0.2-0.48]	0.17 [-0.04-0.53]	0.17 [-0.2-0.75]	0.23 [-0.09-0.86]	0.16 [-0.07-0.57]	0.18 [-0.18-0.65]	0.24 [-0.1-0.76]	0.4 [-0.08-0.93]	0.5 [-0.13-1.34]
North American Monsoon (JAS)											
Precipitation	3.05 [2.24-3.96]	0.13 [-0.08-0.43]	0.07 [-0.27-0.32]	0.02 [-0.32-0.41]	-0.03 [-0.37-0.38]	-0.03 [-0.43-0.52]	0.18 [-0.05-0.44]	0.04 [-0.35-0.39]	-0.1 [-0.51-0.37]	-0.19 [-0.76-0.44]	-0.15 [-0.96-0.57]
Runoff	0.46 [0.09-0.87]	0.03 [-0.04-0.12]	0.03 [-0.07-0.16]	0.02 [-0.1-0.14]	-0.0 [-0.1-0.14]	-0.0 [-0.11-0.14]	0.04 [-0.03-0.15]	-0.0 [-0.19-0.15]	-0.03 [-0.22-0.14]	-0.05 [-0.23-0.19]	-0.06 [-0.29-0.23]
P-E	0.78 [-0.1-1.45]	0.06 [-0.1-0.2]	0.02 [-0.18-0.24]	0.0 [-0.22-0.23]	-0.03 [-0.24-0.2]	-0.04 [-0.31-0.27]	0.09 [-0.06-0.31]	0.01 [-0.22-0.25]	-0.08 [-0.28-0.25]	-0.17 [-0.68-0.25]	-0.18 [-0.72-0.38]
South American Monsoon (DJF)											
Precipitation	8.44 [5.98-10.22]	0.09 [-0.2-0.3]	0.12 [-0.29-0.62]	0.09 [-0.47-0.62]	0.07 [-0.55-0.62]	0.07 [-0.5-0.71]	0.02 [-0.32-0.36]	0.09 [-0.33-0.58]	0.07 [-0.63-0.81]	0.05 [-1.17-0.82]	-0.0 [-1.22-1.19]
Runoff	2.49 [1.11-4.38]	-0.02 [-0.23-0.26]	-0.01 [-0.43-0.53]	0.01 [-0.45-0.46]	-0.03 [-0.49-0.36]	-0.03 [-0.56-0.53]	-0.04 [-0.27-0.28]	-0.01 [-0.41-0.39]	-0.01 [-0.58-0.55]	-0.06 [-0.81-0.24]	-0.04 [-0.85-0.93]
P-E	4.5 [2.83-6.01]	0.04 [-0.23-0.25]	0.08 [-0.26-0.47]	0.04 [-0.43-0.53]	0.04 [-0.5-0.61]	0.02 [-0.45-0.58]	-0.01 [-0.32-0.29]	0.03 [-0.34-0.43]	-0.02 [-0.63-0.62]	-0.02 [-1.03-0.72]	-0.09 [-1.11-0.98]
Australian and Maritime Continent Monsoon (DJF)											
Precipitation	8.63 [6.79-10.7]	0.26 [0.04-0.49]	0.22 [-0.23-0.53]	0.28 [-0.2-0.79]	0.25 [-0.14-0.73]	0.38 [0.0-0.84]	0.15 [-0.09-0.34]	0.24 [-0.36-0.74]	0.5 [-0.1-1.07]	0.65 [-0.08-1.33]	0.9 [0.09-1.76]
Runoff	3.82 [1.78-7.25]	0.2 [-0.01-0.48]	0.23 [-0.11-0.48]	0.29 [-0.11-0.7]	0.24 [-0.13-0.56]	0.35 [-0.03-0.87]	0.12 [-0.06-0.39]	0.29 [-0.08-0.88]	0.49 [0.09-1.25]	0.61 [-0.09-1.05]	0.92 [0.14-1.83]
P-E	4.8 [3.19-6.63]	0.22 [0.03-0.47]	0.13 [-0.23-0.42]	0.2 [-0.16-0.7]	0.2 [-0.14-0.62]	0.27 [-0.09-0.61]	0.12 [-0.1-0.31]	0.16 [-0.31-0.54]	0.38 [-0.05-0.75]	0.54 [-0.08-1.13]	0.69 [0.09-1.27]

West African Monsoon (JJAS)											
Precipitation	5.14 [3.62-7.18]	0.16 [-0.19-0.4]	0.14 [-0.22-0.56]	0.24 [-0.14-0.72]	0.3 [-0.1-0.85]	0.38 [-0.12-1.24]	0.06 [-0.25-0.52]	0.1 [-0.25-0.57]	0.25 [-0.32-0.91]	0.38 [-0.49-1.14]	0.49 [-0.55-1.56]
Runoff	1.43 [0.34-2.57]	0.06 [-0.07-0.22]	0.05 [-0.18-0.27]	0.14 [-0.13-0.54]	0.2 [-0.05-0.7]	0.24 [-0.1-0.8]	-0.01 [-0.2-0.21]	0.03 [-0.25-0.35]	0.1 [-0.25-0.51]	0.25 [-0.28-0.85]	0.3 [-0.33-0.93]
P-E	2.41 [1.05-4.07]	0.08 [-0.2-0.35]	0.1 [-0.2-0.4]	0.2 [-0.11-0.63]	0.23 [-0.11-0.74]	0.36 [-0.06-1.11]	-0.01 [-0.27-0.35]	0.07 [-0.2-0.44]	0.18 [-0.21-0.6]	0.28 [-0.38-0.95]	0.46 [-0.44-1.4]

[START FIGURE 8.22 HERE]

Figure 8.22: Projected regional monsoons precipitation changes. Percentage change in projected seasonal mean precipitation over regional monsoon domains (as defined in Fig 8.11, Section 8.3.2.4 and Annex V) for near-term (2021-2040), mid-term (2041-2060), and long-term (2081-2100) periods based on 24 CMIP6 models and three SSP scenarios (SSP1-2.6, SSP2-4.5 and SSP5-8.5). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.22 HERE]

8.4.2.4.1 South and Southeast Asian Monsoon

In AR5, South and Southeast Asian Monsoon (SAsiaM) precipitation was projected to increase by the end of the 21[st] century but with a weakening of the circulation, with high agreement across the CMIP5 models (Kitoh, 2017; Kitoh et al., 2013; Kulkarni et al., 2020; Menon et al., 2013; Sharmila et al., 2015; Sooraj et al., 2015). Since AR5, most studies have confirmed projected increases in South Asian monsoon precipitation (*high confidence*), while one high-resolution model (35 km in latitude/longitude) projects monsoon precipitation decreases during the 21st century following the RCP4.5 scenario (Krishnan et al., 2016).

Over South Asia, the moisture-bearing monsoon low-level jet is projected to shift northward in CMIP3 and CMIP5 models (Sandeep and Ajayamohan, 2015). Greater warming over the Asian land region compared to the ocean contributes to intensification of the monsoon low-level southwesterly winds and precipitation (Endo et al., 2018), even though the combined effect of upper and lower tropospheric warming makes the Asian monsoon circulation response rather complicated. A high resolution model projection, based on the RCP8.5 scenario, indicates that a northward shift of the low-level jet and associated weakening of the large-scale monsoon circulation can induce a large reduction in the genesis of monsoon low pressure systems by the late 21[st] century (Sandeep et al., 2018). Experiments with constant forcing indicate that at 1.5° and 2°C global warming levels, mean precipitation and monsoon extremes are projected to intensify in summer over India and South Asia (Chevuturi et al., 2018; Lee et al., 2018a) and that a 0.5°C difference would imply a 3% increase of precipitation (Chevuturi et al., 2018). CMIP5 models project an increase in short intense active days and decrease in long active days, with no significant change in the number of break spells for India (Sudeepkumar et al, 2018).

Future monsoon projections from CMIP6 models show an increase of SAsiaM precipitation across all the scenarios and across all the time frames (Figure 8.22) with the maximum increase at the end of the 21[st] century in SSP5-8.5 (Almazroui et al., 2020c; Chen et al., 2020e; Ha et al., 2020; Wang et al., 2020b). Table 8.2 confirms that changes in runoff and P-E over SAsiaM region are positive and largest in the higher emission scenarios considered, as in precipitation. On the other hand, changes in the ensemble mean for all the variables considered in the SSP1-1.9 scenario are negative for both mid and long term periods (Table 8.2). This is also consistently reflected in the spatial map of future precipitation changes (Figure 8.15). Different near-term projections of the SAsiaM may result given the diversity in the future aerosol emission pathways and policies for regulating air pollution (Wilcox et al., 2020). Additionally, near-term projections of SAsiaM precipitation are expected to be constrained by internal variability associated with the PDV (Huang et al., 2020b). CMIP6 models also indicate a lengthening of the summer monsoon over India by the end of the 21[st] century, at least in SSP2-4.5, with considerable inter-model spread in the projected late retreat (Ha et al. et al., 2020).

In summary, consistent with AR5, there is *high confidence* that SAsiaM precipitation is projected to increase during the 21st century in response to continued global warming across the CMIP6 higher emissions scenarios, mostly in the mid and long terms.

8.4.2.4.2 East Asian Monsoon

In AR5, the East Asian monsoon (EAsiaM) was projected to intensify in terms of precipitation, with an earlier onset and longer duration of the summer season. Since AR5, there has been improved understanding of future projected changes in the EAsiaM.

CMIP5 projections indicated a possible intensification of the EAsiaM circulation during the 21st century, in addition to precipitation increase, although there is a lack of consensus on changes in the western North Pacific subtropical high, this is an important feature of the EAsiaM circulation (Kitoh, 2017). Furthermore, the EAsiaM precipitation enhancements in the CMIP5 projections are prominent over the southern part of the Baiu rainband by the late 21st century, with no significant changes in the Meiyu precipitation over central-eastern China (Horinouchi et al., 2019). It was also shown that the Baiu precipitation response in CMIP5 projections is accompanied by a southward retreat of the western North Pacific subtropical high and a southward shift of the East Asian subtropical jet (Horinouchi et al., 2019). According to the high-resolution MRI-AGCM global warming experiments, future summer precipitation could potentially increase on the southern side and decrease on the northern side of the present-day Baiu location in response to downward-motion tendencies which can offset the 'wet-gets-wetter' effect, but is subject to large model uncertainties (Ose, 2019). Future projections of land warming over the Eurasian continent (Endo et al., 2018) and intensified land-sea thermal contrast (Wang et al., 2016c; Tian et al., 2019) can potentially intensify the EAsiaM circulation during the 21st century. However, there are large uncertainties in projected water cycle changes over the region (Endo et al., 2018), mostly in the near-term because of uncertainties in future aerosol emission scenarios (Wilcox et al., 2020), as well as due to the interplay between internal variability and anthropogenic external forcing (Wang et al., 2021).

Interhemispheric mass exchange can act as a bridge connecting Southern Hemisphere circulation with EAsiaM rainfall, however this interhemispheric link is projected to weaken in a future warmer climate as seen from a CCSM4 projection using the RCP8.5 scenario (Yu et al., 2018). A comparison of 1.5°C and 2°C global warming levels reveals how a 0.5°C difference could result in precipitation enhancement over large areas of East Asia (Lee et al., 2018a; Liu et al., 2018b; Chen et al., 2019), with substantial increases in the frequency and intensity of extremes (Chevuturi et al., 2018; Li et al., 2019a). Future monsoon projections from the CMIP6 models show increase of EAsiaM precipitation across all the scenarios (Chen et al., 2020e), though with a large model spread mostly on the long-term and in the higher emission scenarios (Figure 8.22). Considering all the five scenarios used across the report, changes in precipitation, runoff and P-E over the EAsiaM are positive and become larger for highest emission scenarios and for the long term mean, except for the mid-term SSP1-1.9 scenario where the changes are close to zero or even negative (Table 8.2). Additionally, CMIP6 models confirm a projected increased length of the EAsiaM season due to early onset and late retreat (Ha et al., 2020).

In summary, despite the uncertainties in the monsoon circulation response in CMIP5 and CMIP6 models, there is *high confidence* that summer monsoon precipitation over East Asia will increase in the 21st century and *medium confidence* that the monsoon season will be longer.

8.4.2.4.3 West African Monsoon

AR5 concluded that projections of West African monsoon (WAfriM) rainfall are highly uncertain in CMIP3 and CMIP5 models, but still suggest a small delay and intensification in late wet season rains. Studies published since AR5 are broadly consistent with this assessment. CMIP6 models agree on statistically significant projected increases in rainfall in eastern-central Sahel and a decrease in the west for the end of the 21st century (Roehrig et al., 2013; Biasutti, 2019; Monerie et al., 2020). However, the magnitude of WAfriM projected precipitation depends on the convective parameterization used (Hill et al., 2017), and large uncertainties remain in WAfriM projections because of large inter-model spread, particularly over the western Sahel (Roehrig et al., 2013; Biasutti, 2019; Monerie et al., 2020). CMIP6 models show a general increase of WAfriM precipitation across all future scenarios but with a substantial model spread for the SSP5-8.5 scenario (Figure 8.22). This sensitivity arises from the combined and contrasting influences of anthropogenic greenhouse gas and aerosol forcing that affect WAfriM precipitation (particularly over the Sahel) directly and also indirectly through sub-tropical North Atlantic SST changes (Giannini and Kaplan, 2019). The large model spread and associated uncertainties in projected precipitation changes is reflected also in runoff and P-E changes (Table 8.2). Regional climate models (RCMs) ensembles (e.g., Klutse et al.,

2018) agree with CMIP5 projected rainfall trends but some individual models show rainfall declines (e.g., Sylla et al., 2015; Akinsanola et al., 2018), highlighting the existing large uncertainties in RCMs WAfriM rainfall projections.

Changes in seasonality (see also Box 8.2) are projected with a later monsoon onset (*high confidence*) over the Sahel and a late cessation (*medium confidence*), suggesting a delayed wet season as a regional response to global GHG forcing (Biasutti, 2013; Akinsanola and Zhou, 2018; Dunning et al., 2018). Rainfall distribution is projected to be highly variable with a decrease in the number of rainy days in the western Sahel, consistent with an increase in consecutive dry days and a reduction in the number of growing season days (Cook and Vizy, 2012; Diallo et al., 2016). A decrease in the frequency but an increase in the intensity of very wet events is projected to be more pronounced over the Sahel than over Guinean coast, and also under higher emission scenarios (i.e. RCP8.5) (e.g., Sylla et al., 2015; Akinsanola et al., 2018).

In summary, post-AR5 studies and newly available CMIP6 results indicate projected rainfall increases in the eastern-central WAfriM region but decreases in the west (*high confidence*), with a delayed wet season (*medium confidence*). Overall, WAfriM summer precipitation is projected to increase during the 21st century but with larger uncertainty noted under high emission scenarios (*medium confidence*).

8.4.2.4.4 North American Monsoon

AR5 concluded that the North American Monsoon (NAmerM) will *likely* intensify in the future, even though there is *low agreement* among models. AR5 reported *medium confidence* that precipitation associated with the NAmerM will arrive later in the annual cycle and persist longer.

Since AR5, analyses of CMIP5 projections suggest little change in the overall amount of NAmerM precipitation in response to rising global surface temperature. However, significant declines are projected in the early monsoon season and increases in the late monsoon season, suggesting a shift in seasonality toward a delayed monsoon onset and demise (Cook et al., 2013). It is recognised that CMIP5 models are generally too coarsely-resolved to simulate the Gulf of California and the moisture surges associated with the NAmerM (Pascale et al., 2017). Under different RCPs, CMIP5 models tend to project a reduction in NAmerM precipitation but an increase in extreme precipitation events (Torres-Alavez et al., 2014; Bukovsky et al., 2015; Pascale et al., 2019). The almost unchanged or slight decrease in NAmerM total precipitation amount under global warming projections is at odds with paleoclimate records that suggest increased monsoon precipitation under past warm conditions (D'Agostino et al., 2019; Seth et al., 2019). However, there is *low agreement* on how those changes and the mechanisms that drive them are affected under different RCPs since most simulations are model-dependant (Cook and Seager, 2013; Geil et al., 2013; Pascale et al., 2019). Projections from six CMIP6 models show a shortening of the NAmerM under the SSP5-8.5 scenario due to earlier demises (Moon and Ha, 2020). In addition, CMIP6 projections show a decrease in NAmerM precipitation under SSP2-4.5 and SSP5-8.5 scenarios by the end of the 21st century with large inter-model spread (Figure 8.22). This result is also supported by the analysis of 31 CMIP6 models under the SSTP5-8.5 scenario for the 2080-2099 period (Almazroui et al., 2021). Non-linearities and uncertainties in the NAmerM projected changes are valid for many water cycle variables, like precipitation, runoff and P-E (Table 8.2).

In summary, there is *low agreement* on a projected decrease of NAmerM precipitation, however there is *high confidence* in delayed onsets and demises of the summer monsoon.

8.4.2.4.5 South American Monsoon

AR5 reported *medium confidence* that the South American Monsoon (SAmerM) overall precipitation will remain unchanged, and *medium confidence* in projections of extreme precipitation. AR5 also stated *high confidence* in the spatial expansion of the SAmerM, resulting from increased temperature and humidity.

Since AR5, some studies indicate that the SAmerM would experience changes in its seasonal cycle, with delayed monsoon onsets under increasing GHG emissions associated to different RCPs (Fu et al., 2013;

Reboita et al., 2014; Boisier et al., 2015; Pascale et al., 2016; Seth et al., 2019; Sena and Magnusdottir, 2020). In contrast, other studies indicate projected earlier onsets and delayed retreats of the SAmerM under the RCP8.5 scenario based on six CMIP5 models (Jones and Carvalho, 2013). These differences have been linked to the methodology used to determine monsoon timing, and sensitivity to the monsoon domain considered (Correa et al., 2021) (Section 8.3.2.4.5). Recent studies provide further evidence for the projection of delayed SAmerM onsets by the late 21 century (Sena and Magnusdottir, 2020). An analysis of six CMIP6 models under the SSP5-8.5 scenario confirm the projections of delayed SAmerM onsets by the end of the 21^{st} century (Moon and Ha, 2020). In addition, projected changes in the intensity and length of the SAmerM season have been found to be model-dependent (Pascale et al., 2019). The analysis of CMIP5 projections of total monsoon rainfall indicate mixed signals in the Amazon and SAmerM regions (Jones and Carvalho, 2013; Marengo et al., 2014), with some studies suggesting increased summer precipitation in the core SAmerM region (Kitoh et al. et al., 2013; Seth et al., 2013). Dynamical downscaling of CMIP5 projections under the RCP4.5 and RCP8.5 scenarios with the Eta RCM suggests reductions of austral summer precipitation over the SAmerM region throughout the 21^{st} century (Chou et al., 2014). Further analysis using 15 different CMIP6 models for the SSP2-4.5 scenario suggest reductions in total SAmerM rainfall (Wang et al., 2020a). However, other analyses of CMIP6 projections under different SSP scenarios do not report clear changes in the SAmerM precipitation throughout the 21st century (Chen et al., 2020e; Jin et al., 2020), Figure 8.22). Similar uncertainties for all the SSP scenarios used across the report are found for other water cycle variables, including runoff and P-E (Table 8.2). Furthermore, there is disagreement in projected extreme precipitation in the region, with some CMIP5-based studies suggest reductions (Marengo et al., 2014), while others indicate increases based on CMIP5 and CMIP6 models (Kitoh et al., 2013; Sena and Magnusdottir, 2020).

In summary, there is *high confidence* that the SAmerM will experience delayed onsets in association with increases in GHG. However, there is *low agreement* on the projected changes in terms of total precipitation of the South American summer monsoon season.

8.4.2.4.6 Australian and Maritime Continent Monsoon

AR5 concluded that projected changes in Australian and Maritime Continent Monsoon (AusMCM) rainfall and seasonality are uncertain in the CMIP5 models, with some projecting increases and others projecting decreases for the range of emissions scenarios. Models that perform better at simulating present day regional climate project little change or an increase in Australian monsoon rainfall (Jourdain et al., 2013; CSIRO and Bureau of Meteorology, 2015; Brown et al., 2016b). CMIP6 models project increased AusMCM precipitation in the 21^{st} century but with a more robust signal in SSP2-4.5 and SSP5-8.5 rather than in lower emission scenarios (Figure 8.22). A reduced range of CMIP6 rainfall projections but continued disagreement on the sign of change is reported over Australia (Narsey et al., 2020).

The northern and eastern parts of the Maritime Continent have projected increases in rainfall in CMIP5 models (Siew et al., 2014), while there are projected decreases over Java, Sulawesi and southern parts of Borneo and Sumatra. Rainfall changes are correlated with the extent of warming in the western tropical Pacific in CMIP5 models (Brown et al., 2016b) but inter-model differences are also related to modelled large-scale zonal mean precipitation response in both CMIP5 and CMIP6 model ensembles (Narsey et al., 2020). Decomposition of projected rainfall changes indicates that the largest source of model uncertainty is associated with shifts in the spatial pattern of convection (Chadwick et al., 2013; Brown et al., 2016b). Uncertainties in capturing the spatial and temporal features of the Maritime Continent monsoon depend also on the horizontal resolution of coupled climate models (e.g. Jourdain et al., 2013).

The role of anthropogenic aerosol forcing in future projections of the Australian monsoon has been investigated for CMIP5 models (Dey et al., 2019); decreases in anthropogenic aerosol concentrations over the 21^{st} century are expected to produce relatively greater warming in the Northern Hemisphere than Southern Hemisphere, favouring a northward shift of the tropical rain belt (e.g. Rotstayn et al., 2015).

There are some clear projected changes in the rainfall variability and extremes of the Australian monsoon. Rainfall variability in the Australian monsoon domain increases on time scales from daily to decadal in

CMIP5 models (Brown et al., 2017), indicating either more intense wet days or more dry days or both. There is also a projected increase in the intensity of extreme rainfall but a reduction in the frequency of heavy rainfall days for the Australian monsoon (Dey et al., 2019). This is consistent with Moise et al. (2020), who found an increase in Australian monsoon active phase or 'burst' rainfall intensity but a reduction in the number of burst days and events.

Zhang et al. (2013) examined changes in Australian monsoon onset and duration in CMIP3 models and found model agreement on a delay in onset and shortened duration to the north of Australia, but less agreement over the interior of the continent. An updated study of CMIP5 models found similar mean changes with delayed onset and shortened duration, but substantial model disagreement (Zhang et al., 2016).

In summary, CMIP6 projections show an increase of AusMCM precipitation across all emission scenarios. There is strong model agreement on an increase in monsoon precipitation over the Maritime Continent while there is low agreement on the direction of change over northern Australia. There is a projected increase in rainfall variability over northern Australia, with increased intensity of rainfall during the active or 'burst' phase (*medium confidence*).

8.4.2.5 Tropical cyclones

Tropical cyclones (TCs) projections are primarily assessed in section 11.7.1.5. Here, we extend this analysis by assessing the implications of projected changes in tropical cyclones on the water cycle.

AR5 concluded that TC rainfall rate was *likely* to increase through the 21st century. Section 11.7.1.5 assesses that the average tropical cyclone rain-rate is projected to increase with warming (*high confidence*), and peak rain-rates are projected to increase at greater than the Clausius-Clapeyron scaling rate of 7% per 1°C of warming in some regions due to increased low-level moisture convergence (*medium confidence*). The increase in TC rainfall rate is explained by increased TC intensity resulting from increasing SSTs, and increased environmental water vapour (Chauvin et al., 2017; Liu et al., 2019b).

Consistent with the observed poleward migration of tropical cyclone activity (Kossin et al., 2014), in the Southern Hemisphere a larger proportion of storms are projected to decay south of 25°S at the end of the 21st century but with negligible changes in genesis latitude and storm duration for the Australian region (CSIRO and Bureau of Meteorology, 2015)(Sharmila and Walsh, 2018). An analysis of projections for North Pacific islands indicate that the maximum intensity of storms will increase but the number of tropical cyclones will decrease in some places, such as Guam and Kwajalein Atoll in the tropical northwestern Pacific, or remain the same in other regions like near Okinawa (Japan) or Oahu (Hawaii) (Widlansky et al., 2019). TC-induced storm tides affecting landfall in the Pearl River delta over South China are projected to increase by the end of the 21st century (Chen et al., 2020c)

In summary, there is *high confidence* that heavy precipitation associated with tropical cyclones is projected to increase, in response to well-understood processes related to increased low-level moisture convergence and environmental water vapour.

8.4.2.6 Stationary waves

AR5 did not provide an assessment of stationary wave projections as distinct from other related aspects of circulation, such as blocking, modes of variability, and storm tracks. Here we provide a brief assessment of stationary wave projections from the water cycle perspective, with the related circulation aspects considered separately in the following sections.

Several studies based on CMIP5 projections show changes in Northern Hemisphere winter stationary waves that increase precipitation over the west coast of North America and decrease it over the eastern Mediterranean and parts of southwestern North America (Neelin et al., 2013; Seager et al., 2014a, 2014b, 2019b; Simpson et al., 2016; Wills et al., 2019), although the underlying dynamics are not yet fully understood (Seager et al., 2019b; Wills et al., 2019). For the Northern Hemisphere winter global teleconnection pattern, the majority of the models analyzed in (Sandler and Harnik, 2020) project the development of a preferred longitudinal phasing for the pattern, but with strong disagreement among models over the details of the phasing and therefore the associated regional hydrologic impacts.

While the potential role of increasing hydrologic extremes with quasi-resonant stationary waves during Northern Hemisphere summer has received considerable attention (see Section 8.3.2.6), as yet there is no clear evidence in model projections that this variability will increase (Teng and Branstator, 2019). The influence of the Arctic on midlatitude circulation is assessed in Cross-Chapter Box 10.1, which reports that there is *low confidence* in the dominant contribution of Arctic warming compared to other drivers in future projections. Potential changes to the stratospheric polar vortex in CMIP5 models have a substantial influence on tropospheric stationary waves and associated hydrologic impacts in both the Northern (Zappa and Shepherd, 2017) and Southern Hemisphere (Mindlin et al., 2020). CMIP5 models have some important limitations in their representation of stationary waves (Lee and Black, 2013; Simpson et al., 2016; Garfinkel et al., 2020) and this aspect of CMIP6 models has not yet been comprehensively evaluated.

In summary, future changes in stationary waves may have an important influence on both the mean state and variability of the water cycle. Limitations in model representation, dynamical understanding, and the number of targeted studies on the topic currently constrain the assessment of future changes in stationary waves. Based on current knowledge, there is *low confidence* that projected changes in stationary wave activity will contribute to decreases of cold season precipitation over the eastern Mediterranean and increases over the west coast of North America.

8.4.2.7 Atmospheric blocking

In AR5, the increased ability of models to simulate blocking and higher agreement on projections led to an assessment with *medium confidence* that the frequency of Northern and Southern Hemisphere blocking will not increase, but future changes in blocking intensity and persistence were deemed uncertain (AR5 Chapter 14, ES and Box 14.2). Blocking influences precipitation (e.g., Trigo et al., 2004), flooding (e.g., Yamada et al., 2016), drought (e.g., Dong et al., 2018b), snow (e.g., García-Herrera and Barriopedro, 2006), and glacier melt (e.g., Hanna et al., 2013), and so is of broad importance to the water cycle in areas of blocking activity.

Blocking projections are assessed in this Report in Chapter 4 (Section 4.5.1.6), and model performance in simulating blocking is also discussed in Chapter 3 (Section 3.3.3.3). CMIP5 projections suggest a complex response in blocking frequencies with an eastward shift in Northern Hemisphere winter blocking, mid-latitude decreases in boreal summer except in eastern Europe-western Russia, and Southern Hemisphere decreases in the Pacific sector during austral spring and summer. CMIP6 projections (Figure. 4.28) show a notable decrease in blocking activity over Greenland and the North Pacific for the SSP3-7.0 and SSP5-8.5 scenarios. However, the continued large differences among current models as well as the sensitivity to blocking detection methods limits confidence in projected regional changes in blocking (see also Chapter 10, Section 10.3.3.3.1). The influence of blocking on multiple elements of the water cycle means that the uncertainty in blocking projections adds a corresponding layer of uncertainty to water cycle projections.

In summary, and despite recent improvements in the simulation of blocking, there is *limited evidence* in model projections of future changes, except for boreal winter over Greenland and the North Pacific where there is *high confidence* that blocking events are not expected to increase in the SSP3-7.0 and SSP5-8.5 scenarios. As with stationary waves, this adds uncertainty to mid-latitude water cycle projections at the regional scale.

8.4.2.8 Extratropical cyclones, storm tracks and atmospheric rivers

8.4.2.8.1 Extratropical cyclones and storm tracks

AR5 found that extratropical storms were expected to decrease in the Northern Hemisphere, but only by a few percent. Meanwhile, precipitation associated with extratropical storms was projected to increase due to thermodynamic increases in moisture but potentially also due to intensification from increased latent heat release. Latent heating is a strong influence on extratropical storms, so it is plausible that changes in precipitation and associated latent heating could affect extratropical storm intensity and thus precipitation (Zhang et al, 2019).

There is increased evidence that precipitation associated with individual extratropical storms is projected to increase, following thermodynamic drivers with negligible dynamic change (Yettella and Kay, 2017). Comparisons with reanalyses also support the projected increase in thermodynamic precipitation with little dynamic response for precipitation associated with extratropical storms (Li et al., 2014). There is *high confidence* that projected increases in precipitation associated with extratropical storms in over the Northern Hemisphere (Hawcroft et al., 2018; Kodama et al., 2019; Marciano et al., 2015; Michaelis et al., 2017; Pepler et al., 2016; Yettella and Kay, 2017; Zhang and Colle, 2017). A projected decrease in the number of extratropical cyclones over the Northern Hemisphere during the boreal summer in CMIP5 models was reported by (Chang et al., 2016) who related this decrease with a decrease in cloudiness and thus accentuating increased maximum temperatures. However, model spread was quite large, especially over North America, thus there is only *low confidence* in this seasonal signal.

In AR5, the Southern Hemisphere storm track was deemed *likely* to shift poleward, the North Pacific storm track *more likely than not* to shift poleward, while the North Atlantic storm track was *unlikely* to display any discernible changes. There was *low confidence* in regional storm track changes and the associated surface climate impacts, although a weakening of the Mediterranean storm track was a robust response of the models. Since AR5, the Southern Hemisphere mid-latitude storm track is projected to shift poleward and the westerlies are projected to strengthen over Australia (CSIRO and Bureau of Meteorology, 2015). Although thermodynamic effects were considered to be the most important factor in overall projections of increased mid-latitude precipitation, the general poleward shift in cyclogenesis and an enhanced latitudinal displacement of individual cyclones may play a role (Tamarin-Brodsky and Kaspi, 2017).

In the, several factors were identified as relevant to the uncertainties in projections of cyclone intensity, frequency, location of storm tracks and precipitation associated with ETCs. These include horizontal resolution, resolution of the stratosphere, and how changes in the Atlantic meridional overturning circulation (AMOC) were simulated. Since AR5, projections of extratropical cyclones and storm tracks have been examined further, largely confirming previous assessments. In particular, extratropical cyclone precipitation scales with the product of cyclone intensity (as measured by near-surface wind speed) and atmospheric moisture content (Pfahl and Sprenger, 2016). Booth et al. (2018) showed that the fraction of rainfall generated by the convection scheme in simulated extratropical cyclones is highly model- and resolution-dependent, which may be a source of uncertainty regarding their precipitation response to anthropogenic forcings. Also, increased moisture availability may increase the maximum intensity of individual storms while reducing the overall frequency as poleward energy transport becomes more efficient.

The role of temperature trends in influencing storm tracks has been further investigated, both in terms of upper tropospheric tropical warming (Zappa and Shepherd 2017) and lower tropospheric Arctic amplification (Wang et al., 2017), including the direct role of Arctic sea ice loss (Zappa et al., 2018), and the competition between their influences (Shaw et al, 2016). Physical linkages between Arctic amplification and changes in the mid-latitudes are uncertain, as discussed in Chapter 10 (Cross-Chapter Box 10.1). The remote and local SST influence has been further examined by Ciasto et al., (2016), who confirmed sensitivity of the storm tracks to the SST trends generated by the models and suggested that the primary greenhouse gas influence on storm track changes was indirect, acting through the greenhouse gas influence on SSTs. The importance of the stratospheric polar vortex in storm track changes has received more attention (Zappa and Shepherd 2017, Mindlin et al., 2020) and the anticipated recovery of the ozone layer further complicates the role of the stratosphere (Bracegirdle et al, 2020 ; Shaw et al., 2016).

Biases remain in cyclone locations, intensities, cloud features, and precipitation (Catto, 2016, Chang et al., 2016). Uncertainties in projected precipitation changes in many mid-latitude regions can be explained to a large degree by uncertainties in projected storm track or ETC changes. Multiple studies (Chang et al., 2013; Zappa et al., 2015; Chang, 2018) have shown strong relationships between model projected precipitation change in many regions and model projected change in storm track activity near that regions. While front frequency is well represented, frontal precipitation frequency is too high and the intensity is too low (Catto et al., 2015). Some of the bias in storm tracks appears to be related to limitations in model realization of blocking (Zappa et al. 2014). The CMIP6 generation of models has improved representation of storm tracks in both hemispheres (Bracegirdle et al., 2020; Harvey et al., 2020). Simulation of storm tracks and their associated precipitation generally improve with increasing resolution beyond that used in most current climate models (Barcikowska et al., 2018; Jung et al., 2006; Michaelis et al., 2017). In terms of projections, the decreases in cyclone occurrence over the Mediterranean were replicated in a higher resolution model (Raible et al., 2018).

The projected changes in storm tracks and the associated mechanisms have several important implications for water cycle projections. P-E changes in the Mediterranean, California and Chile are directly linked to storm track changes (Zappa et al., 2020). Where the storm tracks are robustly projected to shift (Southern Hemisphere, North Pacific) or weaken (Mediterranean), understanding the physical causes of the related changes in precipitation helps increase confidence in the projections. Understanding the competing influences provides context for why other regions do not exhibit a consistent signal and cautions against regional projections based on individual models. However, model bias and the need for relatively high resolution to reproduce the relevant dynamics is an important overall limit on confidence in current CMIP6 projections.

In summary, there is the *high confidence* that precipitation associated with extratropical storms will increase with global warming in most regions. The Southern Hemisphere storm track will *likely* shift poleward, the North Pacific storm track *more likely than not* will shift poleward, and the North Atlantic storm track is *unlikely* to have a simple poleward shift/ display any discernible changes. There is *low confidence* in regional storm track changes, although a weakening of the Mediterranean storm track is a robust response of the models.

8.4.2.8.2 Atmospheric Rivers

Atmospheric rivers were not assessed in AR5 but are important in the water cycle as they are linked to extreme rainfall, flooding, and changes in terrestrial water storage including melt and ablation of glaciers and snowpack (Section 8.2.2; 8.2.3.2). In a warming world, there is *high confidence* that thermodynamical increases in atmospheric water vapour ensure that atmospheric rivers will become wetter, hence stronger, and longer-lasting (Payne et al., 2020). This is clearly observed in several regional (Gao et al.,, 2015; Gershunov et al., 2019; Hagos et al., 2016; Payne and Magnusdottir, 2015; Ralph and Dettinger, 2011; Warner et al., 2015, Lavers et al., 2013) and in one global study (Espinoza et al., 2018) of atmospheric river activity in CMIP5 model projections. Lavers et al. (2015) indicate that integrated vapour transport under RCP 8.5 and 4.5 could increase, and consequently this thermodynamic response (O'Gorman, 2015) could affect mid-latitudes regions where orographic precipitation is important (Gershunov et al., 2019).

Under continued global warming, more intense moisture transport within atmospheric river events is projected to increase the magnitude of heavy precipitation events on the west coast of the United States (Lavers et al., 2015; Ralph and Dettinger, 2011; Warner and Mass, 2017), in western Europe (Lavers et al., 2015; Ralph et al., 2016; Ramos et al., 2016), and in east Asia (Kamae et al., 2019) (*very likely*). All CMIP5 models analysed agreed under a range of scenarios, except over the Iberian Peninsula (Ramos et al., 2016) where there is only *low confidence* in projected changes. Kamae et al. (2019) reported a 1% increase per degree Celsius warming in the frequency of atmospheric rivers affecting East Asia, but this is strongly affected by SST changes. Emerging evidence of possible regional changes due to dynamical factors are uncertain (Lavers et al., 2013; Gao and others, 2015; Payne and Magnusdottir, 2015). The frequency, magnitude and duration of atmospheric rivers making landfall along the North American west coast are projected to increase (Gershunov et al., 2019). In contrast, Espinoza et al. (2018) suggest that the number of atmospheric river events is projected to slightly decrease globally.

In semi-arid regions where atmospheric rivers have historically been important and precipitation is mainly confined to the cold season, the contribution of atmospheric rivers to annual total precipitation may be expected to grow disproportionately. For example, in California decreases in precipitation frequency are projected as a result of fewer non-atmospheric river storms, while the projected increase in heavy and extreme precipitation events are almost entirely a result of increased atmospheric river activity (Gershunov et al., 2019). Interannual variability in precipitation amounts is projected to increase because of the overall decrease in the frequency of storms but a stronger dependence on extremes (Polade et al., 2014), particularly due to atmospheric rivers (Gershunov et al., 2019), especially where interaction with topography are important (Gershunov et al., 2019; Polade et al., 2014).

In summary, there is *high confidence* that the magnitude and duration of atmospheric rivers are projected to increase in future, leading to increased precipitation. This is projected to increase the intensity of heavy precipitation events on the west coast of the United States and in western Europe (*high confidence*).

8.4.2.9 Modes of climate variability and regional teleconnections

Following on from the assessment of projected changes in modes of climate variability (MoVs) and regional teleconnections (Section 4.5.3), here we assess their consequences for projected water cycle changes.

8.4.2.9.1 Tropical modes

CMIP6 projections indicate that the amplitude of ENSO (Annex IV.2.3) variability will not substantially change during the 21st century (*high confidence*) (Section 4.4.3.2). However, rainfall variability related to ENSO is projected to increase significantly by the second half of the 21st century, regardless of ENSO amplitude (Section 4.5.3.2). Regional precipitation variability associated with ENSO increases due to increases in atmospheric moisture, regardless of changes in ENSO variability itself (Pendergrass et al., 2017). In many regions, the magnitude of the projected changes related to ENSO is small compared with historical interannual variability (Bonfils et al., 2015; Power and Delage, 2018; Perry et al., 2019). Uncertainties in precipitation projections related to ENSO depend on internal variability associated with the mode (Section 8.5.2), hence the need to have relatively large ensembles (~15 members) to adequately estimate uncertainty (Deser et al., 2018; Maher et al., 2018a; Sun et al., 2018a; Zheng et al., 2018).

Even over regions with statistically significant simulated rainfall teleconnections during the historical period, CMIP5 models do not project clear changes (Perry et al., 2019). Nonetheless, CMIP5 models that realistically reproduce Indian summer monsoon rainfall indicate a strengthening of its relationship with ENSO in RCP8.5 projections, though the response is not consistent for different varieties of ENSO events (Roy et al., 2019). Inconsistent changes in the ENSO–Indian summer monsoon relationship in response to global warming in CMIP5 and CMIP6 models may be related to statistical issues rather than dynamical changes (Bódai et al., 2020; Haszpra et al., 2020). Over East Africa during the boreal spring and summer, ENSO teleconnections are projected to become stronger in the future (Endris et al., 2019). Meteorological drought consequences of each strong El Niño are projected to become more severe in the region (Rifai et al., 2019).

Indian Ocean Dipole (IOD, Annex IV.2.4) and Indian Ocean Basin (IOB, Annex IV.2.4) interactions with ENSO are expected to persist in the future (Section 4.5.3.3) but projected changes in the frequency and intensity of events remain uncertain (Hui and Zheng, 2018; Endris et al., 2019; McKenna et al., 2020). Climate extremes such as those associated with the extreme positive IOD event of 2019 are expected to occur more frequently under continued global warming (Cai et al., 2021). Projected changes in IOD teleconnections are linked to model performance in representing the IOD and its remote influence in the present climate, apparently dominated by a positive IOD event-like mean state (Wang et al., 2017a; Huang et al., 2019). Interactions between the IOD and the Indian Ocean mean state, via atmosphere-ocean feedbacks, can affect the behaviour of the IOD (Ng et al., 2018). In the eastern Horn of Africa, OND rainfall is projected to increase because of IOD-ENSO related SST changes in the Indo-Pacific region and associated

Walker circulation changes (Endris et al., 2019).

Sensitivity studies generally project increases in Madden Julian Oscillation (MJO, Annex IV.2.8) precipitation amplitude in a warmer climate, with increases of up to 14% per °C of warming (Arnold et al., 2013, 2015; Caballero and Huber, 2013; Liu and Allan, 2013; Maloney and Xie, 2013; Schubert et al., 2013; Subramanian et al., 2014; Carlson and Caballero, 2016; Pritchard and Yang, 2016; Adames et al., 2017a; Wolding et al., 2017; Haertel, 2018). However, in CMIP5 models with realistic historical MJO behaviour, the precipitation amplitude over the Indo-Pacific warm pool region changes from -4% to +8% per °C in the RCP8.5 scenario relative to the end of the 20th Century (Bui and Maloney, 2018; Maloney et al., 2019). When simulated MJO precipitation amplitude increases with warming, the leading factor for such change is the intensification of the lower tropospheric vertical moisture gradient, that supports stronger vertical moisture advection per unit diabatic heating (Adames et al., 2017b, 2017a; Arnold et al., 2015; Wolding et al., 2017). In idealised simulations with constant carbon dioxide forcing with El Niño-like patterns, the MJO activity penetrates farther east into the central and east Pacific with increased warming (Subramanian et al., 2014; Adames et al., 2017a). Increased MJO convective variability in a warmer climate does not reflect into increased ability of the MJO to force the extratropics (Wolding et al., 2017).

In summary, even though there is *low confidence* in how the tropical MoVs will change in the future (Sections 4.3.3.2 and 4.5.3.3), their regional hydrological consequences, in terms of precipitation, are projected to intensify (*medium confidence*). For example, the ENSO influence on precipitation over the Indo–Pacific sector is projected to strengthen and shift eastward (*medium confidence*). The MJO is projected to intensify in a warmer climate, with increased associated precipitation (*medium confidence*).

8.4.2.9.2 Extra-tropical modes

CMIP6 projections indicate that the Northern Annular Mode (NAM, Annex IV.2.1) is expected to become more positive in winter throughout the 21st century in the SSP3-7.0 and SSP5-8.5 scenarios (Section 4.5.1). In the near-term, the Southern Annular Mode (SAM, Annex IV.2.2) is projected to become less positive than observed during the end of the 20th century during the austral summer in all SSPs scenarios (Section 4.3.3.1).

In the CMIP5 RCP8.5 scenario, increased amplitude and frequency of the North Atlantic Oscillation (NAO, Annex IV.2) during boreal winter (DJF) is associated with higher precipitation in northern Europe and lower precipitation in southern Europe (Tsanis and Tapoglou, 2019). However, large-ensembles analyses show how the NAO leads to significant uncertainty in future changes of regional climate (Section 8.5.2). For example, more than a 85% increase in precipitation is projected over northern Europe, western Russia and much of eastern North America, with similar decreasing resulting in drying over northwestern Africa and regions adjacent to the Mediterranean Sea (Deser et al., 2017).

In the Southern Hemisphere, the positive trend projected for the SAM in the CMIP5 RCP8.5 scenario appears to mitigate the wetting in the mid-to-high latitudes and the drying over the subtropics, but with strong seasonal dependence (Lim et al., 2016). Regional precipitation changes in South America, South Africa, Southern Australia and New Zealand are not well explained by changes in the SAM, but are related to broad-scale changes in north-south temperature gradients associated with enhanced warming of the tropical upper troposphere and strengthening of the stratospheric polar vortex (Mindlin et al., 2020).

In summary, projected changes in the intensity, frequency and phase of extratropical MoVs (see also Section 4.3 and 4.5) may amplify regional changes in precipitation and contribute to an increase in their intra-seasonal and interannual variability (*medium confidence*). Regionally, there are potentially significant precipitation and atmospheric circulation changes associated with changes in extratropical dynamics (*low confidence*).

8.5 What are the limits for projecting water cycle changes?

Understanding the limits to projecting water cycle changes are fundamental for refining climate and hydrological models needed to develop successful climate change adaptation strategies. Regional water cycle projections depend on a range of model-dependent responses (Section 8.5.1) and are also strongly influenced by internal variability, especially in the near term (Section 8.5.2) (Hawkins and Sutton, 2012; Rowell, 2012; Orlowsky and Seneviratne, 2013; Kent et al., 2015; Fatichi et al., 2016; Greve et al., 2018; Chegwidden et al., 2019). CMIP6 models show that different model responses to the same forcing scenario remain the main source of uncertainty for projected changes in regional precipitation (Figure 8.23) (Lehner et al., 2020). Section 8.5.3 assesses the potential for non-linear responses when shifting from low to high global warming levels (James et al., 2017; see also Section 8.4.2.4). While regional uncertainties related to downscaling methods (Section 10.3.3) and impact models (WGII Chapter 4) are not covered here, the added value of regional climate models is briefly discussed (Section 8.5.1.2.2) with a focus on water cycle changes.

[START FIGURE 8.23 HERE]

Figure 8.23: Geographical and zonal mean distribution of the percentage of variance explained by the three sources of uncertainty in CMIP6 projections of 20-year mean precipitation changes in 2021–2040 (top), 2041–2060 (middle) and 2081–2100 (bottom) relative to the 1995–2014 base period: internal climate variability (left), model response uncertainty (middle) and scenario uncertainty (right, considering four plausible concentration scenarios: SSP1-2.6, SSP2-4.5, SSP3-7.0 and SSP5-8.5). Percentage numbers give the area-weighted global average value for each map. Right panels show the zonal mean fractions over both land and sea (solid lines) and over land only (dashed line). The figure was adapted from Fig.4a in (Lehner et al., 2020). The relative contributions of internal variability, models and emission scenarios to the total uncertainty depend on both region and time horizon. The scenario uncertainty is relatively low in near and mid-term time horizons while it increases in the long-term mostly over the high-latitudes. The model response uncertainty is the most influential factor across all time horizons. Internal variability also plays a key role in the near-term, especially in the subtropics. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.23 HERE]

8.5.1 Model uncertainties of relevance for the water cycle

Model response uncertainty is typically estimated as the inter-model spread (range) projected by a set of climate models for a given emission scenario. It is best estimated at the end of a high-emission scenario when internal variability has a limited contribution to total uncertainty (Figure 8.23). Even for aggregated quantities, like decadal-mean precipitation averaged over relatively large domains, model response uncertainty is substantial and can exceed scenario uncertainty (Hawkins and Sutton, 2011; Lehner et al., 2020). This can also be true for other water cycle variables such as soil moisture, runoff and streamflow at the regional scale, either derived directly from global climate models (GCMs) or produced by "offline" using global hydrological models (GHMs) driven by the same GCMs (Orlowsky and Seneviratne, 2013; Giuntoli et al., 2015, 2018; Chegwidden et al., 2019). Although some of the model response uncertainty is related to climatological biases (Grose et al., 2017; Li et al., 2017b; Lehner et al., 2019; Samanta et al., 2019), model biases are not the only way to assess the reliability of climate projections (cf. Box 4.1). Therefore, our focus here is on the representation of key processes that are not completely resolved in current-generation GCMs (Section 8.5.1.1) and on the model improvements associated with increased horizontal resolution (Section 8.5.1.2).

8.5.1.1 Fitness-for-purpose and poorly constrained key processes

AR5 Chapter 7 recognized that the simulation of clouds and precipitation remains challenging for state-of-the-art GCMs. Model development and evaluation have continued since AR5, with a particular emphasis on the representation of new model components, like interactive vegetation, aerosols and biogeochemical cycles. For example, the comparison of simulated tropical precipitation across three successive generations of CMIP models (including CMIP6) indicates overall little improvement for the summer monsoons, the double-ITCZ bias, the diurnal cycle and the frequency of precipitation (Fiedler et al., 2020). Some of these issues are related to inherent model limitations in three specific areas: atmospheric convection, cloud-aerosol interactions and land surface processes (ocean and cryosphere-related processes are addressed in Chapter 9). These limitations do not weaken the overall progress made in the large-scale simulation of present-day climate (FAQ 3.3, Section 3.3.2.2), even though the improvement of CMIP6 compared with CMIP5 models is limited (Figure 3.12) and is generally less systematic or obvious at the regional scale (e.g., Gusain et al., 2020; Monerie et al., 2020; Oudar et al., 2020). Instead, they call for a careful interpretation of hydrological projections with the full range of plausible outcomes, rather than only considering the most likely scenarios (Sutton, 2018, 2019).

8.5.1.1.1 Atmospheric convection

Moist convection is fundamental to the water cycle through its vertical transport of momentum, heat, and moisture across the atmosphere. It is particularly active in the tropics where it contributes to more than half of annual precipitation and to the development of severe weather events. Given limitations in computing resources, the current-generation GCMs cannot yet represent small-scale cloud processes and consequently shallow and deep convection is determined by sub-grid-scale parameterizations. While such parameterizations can be evaluated against field observations (e.g., Abdel-Lathif et al., 2018), it remains challenging to estimate convective entrainment that is valid for both shallow and deep convection (Zhang et al., 2016a). Comparisons between regional projections with explicit versus parameterized convection also highlight the limitations of parameterized convection for assessing climate change (Kendon et al., 2019; Jackson et al., 2020).

Atmospheric convection is particularly important for a realistic simulation of tropical precipitation intensities (Pendergrass and Hartmann, 2014a; Kendon et al., 2019). Many CMIP5 models produce rainfall at water vapour amounts lower than in observations (Takahashi, 2018), as well as too light and too frequent precipitation events (Sun et al., 2015; Trenberth et al., 2017). Such biases can be explained by a lack of convective inhibition (Rochetin et al., 2014a, 2014b) and by too much convective and too little non-convective precipitation (Chen and Dai, 2019). Tropical convection controls the amount of precipitable water simulated over the equatorial Indian Ocean, which has been identified as a key metric for differentiating model skill in simulating South Asian monsoon precipitation (Hagos et al., 2019). Many models have difficulty in adequately simulating the diurnal cycle of precipitation over land (Couvreux et al., 2015), the rainfall intensity distribution associated with the West African monsoon (Roehrig et al., 2013), and the intensity of tropical cyclones (Sections 10.3.3.4 and 11.7.1.3), phenomena for which atmospheric convection also plays a key role.

Since AR5, there have been improvements in the representation of convective clouds and related precipitation in GCMs. For instance, the drizzle issue (too light and too frequent rainfall events) has led to modifications in the deep convection triggering scheme (Rochetin et al., 2014b; Han et al., 2017; Wu et al., 2018a; Xie et al., 2018). Although high-resolution studies have highlighted these limitations, most GCMs still rely on a convective available potential energy (CAPE) closure which has been adapted to various cloud regimes (Bechtold et al., 2014; Han et al., 2017; Walters et al., 2017) or evaluated against convection-permitting models (CPMs) (Chen et al., 2020b). To increase the sensitivity of convection to tropospheric humidity, several models now include a representation of deep convective entrainment dependent on relative humidity (Bechtold et al., 2008; Han et al., 2017; Walters et al., 2017; Zhao et al., 2018c). Other efforts have focused on the improvement of shallow convection and low-level cloudiness due to their major contribution to uncertainty in climate sensitivity (Section 7.4.2.4). A cloud regime-based study however highlights an apparent disconnection between cloud and precipitation processes in GCMs (Tan et al., 2018), suggesting that a good representation of clouds does not lead to systematic improvement in simulated precipitation. A global simulation in which the parameterized convection is switched off shows a strong influence of parameterized convection on daily precipitation extremes (Maher et al., 2018b). Regional simulations at a 25km resolution suggest that an explicit deep convection can be beneficial even at such a relatively coarse resolution (Vergara-Temprado et al., 2020). Perturbed physics ensembles (PPE, Section 1.4.4) make it possible to identify parameters in the convection scheme that are most important in determining future precipitation changes (Bernstein and Neelin, 2016).

Since AR5, spatial aggregation of tropical convection has also received growing attention in both observational (Holloway et al., 2017) and modelling studies (Muller and Bony, 2015; Wing et al., 2017; Tan et al., 2018). The changing degree of convective organization was highlighted as a key mechanism for dynamic changes in extreme precipitation (Pendergrass, 2020a). Yet, convective parameterizations do not represent all aspects of mesoscale convective systems (Hourdin et al., 2013; Park et al., 2019). This is related to the complexity of mechanisms involved from synoptic to mesoscale dynamics, which are only partially resolved by models. Cloud-resolving models (CRMs, Section 8.5.1.2.2) represent a useful benchmark for improving the parameterization of mesoscale convective systems. Machine learning can also be used to parameterize moist convection after training the model with a conventional or a super parameterization scheme (Gentine et al., 2018; O'Gorman and Dwyer, 2018), but has not yet been used in the CMIP framework.

While some global modelling centres have reported progress in their parameterization of convection and in their simulation of seasonal, daily and sub-daily precipitation (e.g., Danabasoglu et al., 2020; Roehrig et al., 2020), CMIP6 models as a whole only show limited improvements in their simulation of the tropical precipitation climatology compared to CMIP5 (Fiedler et al., 2020; see also Figure 3.10). For instance, the double-ITCZ syndrome is still prominent (Tian and Dong, 2020) despite being reduced in some models (e.g., Qin and Lin, 2018). This systematic bias was shown to arise from atmospheric processes including cloud feedbacks (Tian, 2015; Dixit et al., 2018; Talib et al., 2018) and the SST threshold at which deep convection occurs in the tropics (Oueslati and Bellon, 2015; Xiang et al., 2017; Adam et al., 2018). Such biases can also arise from a too weak sensitivity of seasonal tropical precipitation to local SSTs compared with observations (Good et al., 2020). These biases are large enough to alter forced precipitation changes, and consequently limit our confidence in projected precipitation changes (Samanta et al., 2019; Aadhar and Mishra, 2020). Observational constraints can be used to narrow model response uncertainties (DeAngelis et al., 2015; Li et al., 2017b; Ham et al., 2018; Watanabe et al., 2018), although there is still no consensus that model selection or weighting is a reliable alternative to the 'one-model-one-vote' approach used in Section 8.4 (Box 4.1). The detrimental influence of model errors can also be mitigated by focusing on phenomena or events (Polson and Hegerl, 2017; Weller et al., 2017), implementing bias adjustment techniques (Section 10.2.3.2), or adopting a non-probabilistic storyline approach (Zappa and Shepherd, 2017).

In summary, since AR5 empirical convective parameterization schemes and associated precipitation biases have improved in some but not all global climate models. There is still *low confidence* in their ability to accurately simulate the spatio-temporal features of present-day precipitation, especially in the tropics where a double-ITCZ bias is still apparent in many models. While such biases limit the reliability of precipitation projections in some cases, there is currently only *medium confidence* that model selection or weighting is a better alternative to the one-model-one-vote approach (Box 4.1). Improved water cycle projections can be achieved by focusing on phenomena or weather events, such as a thermodynamic intensification of convective events (*high confidence*, Section 8.2.2.1), however accurate quantitative estimates are currently hampered by complex, model-dependent dynamical responses (Section 8.2.2.2).

8.5.1.1.2 Aerosol microphysical effects on clouds and precipitation

In AR5 Chapter 7, there was *low confidence* in the representation of cloud–aerosol interactions in climate models. Despite progresses in this field since AR5, cloud–aerosol interactions remain a major obstacle to understanding climate and severe weather (Varble, 2018). High aerosol concentrations have been observed to suppress rain in water clouds (Campos Braga et al., 2017; Fan et al., 2020). However, such aerosol effects are muted in GCMs, which tend to produce precipitation from shallow clouds too frequently at the expense of rain intensity (Suzuki et al., 2015; Jing et al., 2017). This arises from incomplete knowledge of how clouds adjust to aerosol primary effects such as cloud condensation nuclei (CCN). The adjustment occurs mainly as a dynamic response to the impacts of CCN on cloud drop size and number concentrations on precipitation-forming processes (Rosenfeld et al., 2008; Goren and Rosenfeld, 2014; Koren et al., 2014; Camponogara et al., 2018). Uncertainties are large for deep clouds, as their processes are much more complex and include also the impacts of aerosols on ice precipitation processes. Aerosols can substantially invigorate (Rosenfeld et al., 2008; Koren et al., 2014; Fan et al., 2018) and electrify (Thornton et al., 2017;

Wang et al., 2018b) deep tropical convective clouds. High-resolution atmospheric simulations suggest that high aerosol concentrations can increase environmental humidity by producing clouds that mix more condensed water into the surrounding air, which in turn favours large-scale ascent and strong convective events (Abbott and Cronin, 2021). Further assessment of uncertainties in aerosol-cloud interactions for shallow water clouds is provided in Section 7.3.3.2.

A major challenge in representing convective clouds and related precipitation events in GCMs is a lack of sophisticated cloud microphysics in convective parameterization schemes (e.g., Fan et al., 2016). Most of these schemes only include simple microphysical treatments, such as direct partition between cloud condensation and precipitation, and do not include advanced treatment of conversion among different types of hydrometeors. As such these schemes are unable to simulate microphysical cloud and precipitation responses to aerosol-related perturbations in cloud droplet concentration and ice crystals (see Box 8.1), or perturbations in thermodynamical states from global warming. Efforts have been made to include more advanced cloud microphysical treatment in cumulus parameterizations (Song and Zhang, 2011; Grell and Freitas, 2014; Berg et al., 2015) or to use explicit cloud microphysics schemes in climate models with a 'super parameterization' (Wang et al., 2015), which have been shown to improve the performance in simulating cloud properties and precipitation. However, few of these improvements have been incorporated into CMIP6 climate models so the projected precipitation response to anthropogenic perturbation may still be hindered by the inadequate microphysical treatment in cumulus parameterization (Smith et al., 2020).

In summary, there is still *low confidence* in the simulated influence of the aerosol microphysical effects on future precipitation changes.

8.5.1.1.3 Land surface processes

Land surface processes determine the partitioning of net surface radiation into sensible, latent and ground heat fluxes, the partitioning of precipitation into evapotranspiration and runoff, and the net terrestrial carbon flux at the Earth's surface. They are relevant for simulating the terrestrial water cycle responses to climate change, as well as the response to land use change (FAQ 8.1). Even basic land surface properties such as albedo (Terray et al., 2018) or the ratio of transpiration to total evaporation (Chang et al., 2018) still need to be improved in state-of-the-art coupled GCMs. Runoff sensitivities are also not well constrained in these models, which display a large spread for the present-day climate, influencing simulated changes under global warming (Lehner et al., 2019). Earth System Models (ESMs) incorporate some combined biophysical and biogeochemical processes to a limited extent, and many relevant processes about how plants and soils interactively respond to climate changes are yet to be considered (e.g., Liu et al., 2020). Consequently, land surface processes and their atmospheric coupling contribute to the range in water cycle projections (Jia et al., 2020).

Since AR5, development of new and existing processes in land surface models (LSMs) have been evaluated. These include soil freezing and permafrost (Vergnes et al., 2014; Chadburn et al., 2015; Yang et al., 2018a; Gao et al., 2019), soil and snow hydrology (Brunke et al., 2016; Decharme et al., 2016), glaciers (Shannon et al., 2019), surface waters and rivers (Decharme et al., 2012), as well as vegetation (Bartlett and Verseghy, 2015; Betts et al., 2015; Knauer et al., 2015; Tang et al., 2015) and the representation of hydraulic gradients throughout the soil-plant-atmosphere continuum (Bonan et al., 2014). Such land surface model developments have led to significant improvements in global off-line hydrological simulations driven by observed atmospheric forcings (e.g., Li et al., 2017a; Decharme et al., 2019).

Progress in the representation of land surface heterogeneity has been made, in the form of improved mapping of root zone storage capacity (Wang-Erlandsson et al., 2016), improved vegetation stand, disturbance and fire dynamics (Li et al., 2013a; Fisher et al., 2018; Haverd et al., 2018; Yue et al., 2018; Zou et al., 2019), better representation of urban surfaces (Box 10.3), and the explicit representation of inland water bodies (Gu et al., 2015; Verseghy and MacKay, 2017). The representation of realistic snow and vegetation cover significantly affects the simulation of the land surface energy and water budgets at multiple time-scales (Loranty et al., 2014; Bartlett and Verseghy, 2015; Thackeray et al., 2015; Qiu et al., 2016; Thackeray and Fletcher, 2016; Wang et al., 2016a; Alessandri et al., 2017). Groundwater remains inadequately represented in many models, which limits our current understanding of the two-way interactions between groundwater and the rest of the hydrologic cycle (Taylor et al., 2013b; Leng et al., 2014; Vergnes et al., 2014; Pokhrel et al., 2015; Maxwell and Condon, 2016; Collins, 2017; Scanlon et al., 2018; Condon et al., 2020). Land management exerts an increasing influence on the water cycle (Abbott et al., 2019) whose representation in the current-generation climate models is generally incomplete (Section 10.3.3.7.2).

Aside from land surface models (LSMs), global hydrological models (GHMs) have been further developed for off-line simulations of the hydrological impacts of both climate change and water management (Cisneros et al., 2014; Schewe et al., 2014; Döll et al., 2016, 2018b, Pokhrel et al., 2016, 2017; Veldkamp et al., 2018). GHMs can equal or outweigh the contribution of GCMs to uncertainties in hydrological projections at the regional scale (Giuntoli et al., 2015). Historical GHM simulations are currently not sufficient to improve regional water cycle projections, due to modelling uncertainties in both the driving GCMs and land surface hydrology (Pechlivanidis et al., 2017; Samaniego et al., 2017; Hattermann et al., 2018; Krysanova et al., 2018). Biophysical vegetation processes are still not accounted for in many GHMs, which may lead to inadequate projections of terrestrial runoff and water resources. However, hydrological models that do simulate these effects often disagree (Prudhomme et al., 2014), so do not necessarily provide the added value of a more sophisticated representation of vegetation processes and land surface conditions (Döll et al., 2016).

Since AR5, there has been increasing recognition of the need to better understand the role of land–atmosphere coupling and related feedbacks (Joetzjer et al., 2014; Berg et al., 2016; Catalano et al., 2016; Berg and Sheffield, 2018a; Santanello et al., 2018). This has led to the development of dedicated field campaigns (Song et al., 2016; Phillips et al., 2017; Dirmeyer et al., 2018), remotely sensed observations (Ferguson and Wood, 2011; Roundy and Santanello, 2017), and tailored diagnostics (Tawfik et al., 2015a, 2015b, Miralles et al., 2016, 2019; Dirmeyer and Halder, 2017). Dynamic vegetation models have been introduced in global ESMs but they need further evaluation (Medlyn et al., 2015; Prentice et al., 2015; Cantú et al., 2018; Franks et al., 2018) to provide valuable information on potential vegetation feedbacks. Plant migration and mortality, increased disturbances from wild fires, insects and extreme events, interactive nitrogen cycle, or the impact of increased levels of tropospheric ozone are often ignored or poorly represented in the current-generation of ESMs (Bonan and Doney, 2018; Fisher et al., 2018).

The physiological response of plants to increasing atmospheric CO_2 is generally accounted for, but only using empirical models of stomatal conductance that are characterized by a single critical parameter of intrinsic water-use efficiency (Franks et al., 2017, 2018). This reflects a lack of structural diversity and caution about the consensus of the photosynthesis response to increasing CO_2 (Knauer et al., 2015; Huang et al., 2016), which has implications for the ability of the current-generation models to account for uncertainty in future evapotranspiration changes. Most CMIP5 models underestimate the ratio of plant transpiration to total terrestrial evapotranspiration, which may suggest that they also underestimate the impact of plant physiology on the water cycle (Lian et al., 2018). Plant hydraulics are not explicitly considered in many land surface models, which may lead to an underestimation of the influence of the increasing atmospheric moisture stress on plant transpiration under climate change (Massmann et al., 2019; Grossiord et al., 2020; Liu et al., 2020c). Most ESMs underestimate the water use efficiency measured at many sites and, consequently overestimate the ratio of evapotranspiration to precipitation (Li et al., 2018a).

In summary, since AR5 substantial advances have been made in the representation of land surface processes in current-generation Earth System Models (ESMs). Off-line hydrological models allow the application of bias-adjusted atmospheric forcings, but there is *low confidence* of an improved response compared to coupled climate models, given their inherent limitations (Box 10.2). While improvements in the representation of complex land surface feedbacks relevant to the water cycle are needed, there is currently *low confidence* that they will substantially improve of water cycle projections.

8.5.1.2 Added value of increased horizontal model resolution

Coarse spatial resolution of climate models has often been considered a key limitation in global climate projections (Di Luca et al., 2015; Roberts et al., 2018). Proposed and tested solutions include a uniform or regional increase in the resolution of GCMs, or the use of regional climate models (RCMs). The increase in computing resources has also led to the development of convection-permitting models (Prein et al., 2015), which have been integrated over larger domains, but are still unsuitable for CMIP simulations. Statistical downscaling tools are also widely used to generate fine-scale regional climate information necessary for climate impacts and adaptation studies. A comprehensive assessment of the added value of increased spatial resolution and of the benefits and shortcomings of statistical downscaling tools are addressed in Chapter 10 (Section 10.3.3).

8.5.1.2.1 High-resolution global climate models

Since AR5, horizontal resolution has increased in most global climate models, which has led to several improvements in the simulation of the water cycle (see also Section 10.3.1.1), not only in areas with steep or complex orography, but also over the tropical oceans and within the North Pacific and North Atlantic storm tracks (Piazza et al., 2016; Roberts et al., 2018; Bui et al., 2019; Chen and Dai, 2019; Vannière et al., 2019). Yet, the added value of higher resolution global climate models is not systematic (Johnson et al., 2016; Ogata et al., 2017; Huang et al., 2018a; Mahajan et al., 2018; Vannière et al., 2019) and needs careful assessment (Haarsma et al., 2016; Caldwell et al., 2019). Several AGCM studies suggest that increased spatial resolution leads to better simulation of the atmospheric moisture transport from ocean to land, the geographical distribution of annual mean precipitation (Demory et al., 2014), and the frequency distribution of daily precipitation intensities (Zhang et al., 2016c; Chen and Dai, 2019) including extremes in many (Jacob et al., 2014; Westra et al., 2014) but not all cases (Bador et al., 2020).

Part of the improvement in simulated precipitation accuracy is related to improved simulation of the frequency and/or mean intensity of tropical (Roberts et al., 2015; Walsh et al., 2015) and extratropical (Hawcroft et al., 2016) cyclones. Idealized regional experiments also show that the North Atlantic storm-track response to global warming can be amplified in higher resolution models (Willison et al., 2015). Increased atmospheric horizontal resolution can be also important for simulating Northern Hemisphere blockings (Davini et al., 2017; Schiemann et al., 2017) and synoptic features of the East Asian summer monsoon (Yao et al., 2017; Kusunoki, 2018). Variable resolution based on grid stretching may be a valuable alternative for simulating regional phenomena like monsoons (P Sabin et al., 2013; Krishnan et al., 2016) or tropical cyclones (Harris et al., 2016; Chauvin et al., 2017), while avoiding inconsistencies in the forcings or physics that can be found in RCMs driven by GCMs (Boé et al., 2020; Tapiador et al., 2020).

Increasing horizontal model resolution in CMIP5 and CMIP6 models leads to a systematic increase in global mean precipitation, enhanced moisture advection to land in close connection with increased orographic precipitation, and a partial reduction of the long-standing double ITCZ bias (Demory et al., 2014; Caldwell et al., 2019; Vannière et al., 2019). Recent studies based on HighResMIP simulations (Haarsma et al., 2016) confirm the added value of increased horizontal resolution (at least 50 km in the atmosphere and 25 km in the ocean) for the simulation of tropical (Roberts et al., 2020) and extratropical cyclones (Priestley et al., 2020). CMIP6 model biases in annual mean precipitation are only slightly reduced at higher resolution (Figure 3.10).

High resolution representation of the land surface is also important for simulating many features of the terrestrial water cycle, such as orographic precipitation, snow, runoff and streamflow in complex topography areas (Zhao and Li, 2015). However, the added value may be easier to assess in off-line rather than online land surface simulations (Döll et al., 2016) given the possible use of bias-corrected atmospheric forcings. Off-line high-resolution GHMs are routinely used to monitor water resources or to assess the hydrological impacts of bias-adjusted global climate projections (Davie et al., 2013; Huang et al., 2017, 2018b). Yet, the development and calibration of 'hyper-resolution' hydrological models, with grid cells of typically 100m to 1km, raises a number of issues given the lack of comprehensive surface or subsurface information (Bierkens et al., 2015) and the lack of coupling with the atmosphere (Berg and Sheffield, 2018a).

In summary, there is *high confidence* that increasing horizontal resolution in GCMs can reduce a number of systematic model errors of relevance for the water cycle, including synoptic circulation and the accuracy of daily precipitation projections. High-resolution GCMs and GHMs provide improved representation of land surfaces, including topography, vegetation and land use change, which are required to accurately simulate changes in the terrestrial water cycle. However, there is *low confidence* that the higher horizontal resolution simulations currently available provide more accurate projections of the large-scale features of the water cycle.

8.5.1.2.2 Regional Climate Models and Convective Permitting Models

Regional Climate Models (RCMs) are used to dynamically downscale global model simulations for a particular region (usually at a spatial resolution in the order of 10 to 50 km) (see Section 10.3.3). AR5 reported that RCMs are useful for regions with variable topography and for small-scale phenomena, however inherit biases from their driving GCMs and thus may lack physical consistency with them. Since AR5, the application of RCMs has largely increased due to international model inter-comparison projects such as CLARIS-LPB (Sánchez et al., 2015). Many studies have focused on present-day climatological precipitation, showing with *high confidence* improvements in its monthly to seasonal accumulation and spatial distribution (Dosio et al., 2015; Giorgi et al., 2016; Bozkurt et al., 2019; Falco et al., 2019; Di Virgilio et al., 2020), although the modelling of precipitation remains the 'Achilles heel' of both GCMs and RCMs and should be considered cautiously when informing regional climate change adaptation strategies (Tapiador et al., 2019a).

Regional Convective Permitting Models (CPMs), typically run at a resolution less than 10 km, have been implemented over increasingly large domains. Compared to models with parameterized convection (Section 8.5.1.1.1), they generally show improved simulation of key features of the water cycle such as orographic precipitation, sea-breeze dynamics, the diurnal cycle in precipitation, soil-moisture precipitation feedbacks, daily precipitation persistence, sub-daily to daily precipitation intensities and related extremes (Birch et al., 2015; Prein et al., 2015; Willetts et al., 2017; Kendon et al., 2017; Leutwyler et al., 2017; Hohenegger and Stevens, 2018; Berthou et al., 2019b; Takahashi and Polcher, 2019; Fumière et al., 2020; Scaff et al., 2020; Caillaud et al., 2021; see also Section 8.2.3.2). A growing number of studies have also assessed the potential added value of using CPMs for regional climate projections (Ban et al., 2015; Giorgi et al., 2016; Fosser et al., 2017; Kendon et al., 2017, 2019; Liu et al., 2017; Rasmussen et al., 2017; Lenderink et al., 2019; see also Atlas 5.6.3). Although projected changes in rainfall occurrence in CPMs are broadly and qualitatively consistent with the results of GCMs and RCMs (Kendon et al., 2017), there is a tendency towards stronger changes in both wet and dry extremes (Berthou et al., 2019a; Kendon et al., 2019; Lenderink et al., 2019; Finney et al., 2020a). While both GCMs and RCMs project an overall decrease in summer precipitation over the Alps, RCMs simulate an increase over the high Alpine elevations that is not present in the global simulations (Giorgi et al., 2016).

Recent studies based on both GCMs and CPMs indicate that both CAPE and convective inhibition will increase in a warmer climate (Chen et al., 2020; see also Section 8.2.3.2), consistent with a shift from moderate to less frequent but stronger convective events (Rasmussen et al., 2017). If underestimated by models with parameterized convection, such a mechanism could explain the underestimation of both projected increase in precipitation extremes (Borodina et al., 2017; Yin et al., 2018) and land surface drying (Douville and Plazzotta, 2017) in the extratropics. CMIP5 models with a larger increase in extreme precipitation also exhibit larger declines or smaller increases in light-moderate events (Thackeray et al., 2018).

In summary, there is *high confidence* that dynamical downscaling using limited area models adds value in simulating precipitation and related water cycle processes at the regional scale, especially in complex orography areas (see also Section 10.3.3.5.1). There is *high confidence* that the explicit simulation of atmospheric convection can improve the representation of weather phenomena, including the life cycle of convective storms and related precipitation extremes. Even with an improved simulation of small-scale processes, there is only *medium confidence* that there will be an improvement in RCM-based water cycle projections as they rely on GCM boundary conditions.

8.5.2 Role of internal variability and volcanic forcing

Beyond modelling uncertainties, internal variability and unpredictable natural forcings may also lower the degree of confidence in projected water cycle changes, especially in the near-term (2021-2040) and regional-scale projections (Hawkins and Sutton, 2011; Kent et al., 2015; Thompson et al., 2015; Fatichi et al., 2016; McKinnon and Deser, 2018; Chen and Brissette, 2019; Lehner et al., 2020). Although there is *low confidence* that the main modes of climate variability (Annex IV) are altered in a warmer climate (Sections 4.4.3 and 4.5.3), increasing contrast between wet and dry weather regimes (Section 8.2.2.1) will amplify their influence on water cycle variability (Section 8.4.2.9) and therefore contribute to uncertainties in near-term precipitation changes (Figure 8.23). The role of internal variability as source of uncertainties in regional climate projections is assessed in Section 10.3.4.3. Here we assess the role of internal variability in influencing water cycle projections using paleoclimate reconstructions, preindustrial model simulations, and large single model ensembles (Section 8.5.2.1). Implications for the predictability of near-term water cycle changes are specifically assessed, as they show significant but model-dependent regional hydrological fingerprints over land (Section 8.5.2.2). The role of volcanic eruptions is also briefly assessed in terms of consequences and uncertainties in water cycle projections (Section 8.5.2.3).

8.5.2.1 Quantification of water cycle internal variability

Estimating internal variability is an important challenge in the assessment of human-induced changes in the water cycle since its magnitude and range of variability can exceed the anthropogenic signal, at least at the regional scale and for near-term projections or low-emission scenarios (Deser et al., 2012; Shepherd, 2014; Xie et al., 2015; Sarojini et al., 2016; Dai and Bloecker, 2019; Lehner et al., 2020) (Section 4.4.1.4, 8.4.2.9). Underestimating internal variability in models may result in the overestimation of anthropogenic climate change because the 'noise' in the signal-to-noise ratio is underestimated (Knutson and Zeng et al., 2018). There is *medium confidence* that this underestimation affects global water cycle projections, for instance, in terms of drought persistence and severity in the southwestern United States, eastern Australia, southern Africa, the Mediterranean, the southern Amazon basin and China (Ault et al., 2014; Cook et al., 2018; Gu et al., 2018). In CMIP6 models, the uncertainty in future projections of 20-year mean precipitation changes attributable to internal variability ranges from 41% in the near-term (2021-2040) to 5% in the long-term (2081-2100) (Figure 8.23). For decadal-mean precipitation changes, the relative contribution of internal variability is even larger when using large ensembles (Lehner et al., 2020).

Over the 20th century, CMIP5 models show a realistic magnitude of decadal precipitation variability, if not a slight overestimation in some regions (Knutson and Zeng, 2018). However, the relatively short and human-influenced instrumental record limits our ability to quantify the magnitude of internal variability in the water cycle, particularly over long timescales (decadal and beyond). Global extended reanalyses (Section 1.5.2) have been used to derive long-term variability in the regional water cycle components (Caillouet et al., 2017), merged with historical meteorological and hydrological local observations (Bonnet et al., 2017; Devers et al., 2020). Specific assessment of these types of methodology and related uncertainties is provided in Chapter 10 (Sections 10.2 and 10.3). Paleoclimate archives (tree rings, corals, ice core, speleothems, lake and ocean sediments) provide extended reconstructions of key water cycle metrics and large-scale circulation features. Some studies have suggested that CMIP5 models underestimate internal variability at decadal and longer timescales, and therefore may be missing important processes in the climate system (Ault et al., 2012, 2013; Bunde et al., 2013; Franke et al., 2013; Cheung et al., 2017; Hope et al., 2017; Kravtsov, 2017; Cassou et al., 2018). However, recent assessments using paleoclimate records have found that CMIP5 models are able to reproduce decadal-to-centennial variability, including the severity, persistence and spatial extent of megadroughts (Coats et al., 2015; Stevenson et al., 2015; PAGES Hydro2K Consortium, 2017), once signal reddening (autocorrelation) in proxy archives is accounted for (Dee et al., 2017; PAGES Hydro2K Consortium, 2017). Implementation of proxy system models, i.e., functions that transform model variables into proxy units, has reduced model-proxy disagreement, although some differences in the magnitude of internal variability remain, particularly at centennial timescales (Dee et al., 2017; Parsons et al., 2017). It is unclear whether remaining discrepancies represent limitations of the climate models, or limitations of the proxy system models. Therefore, there is *medium to high confidence* (i.e., depending on the region) that climate models do not underestimate water cycle internal variability.

The mechanisms driving internal variability in the water cycle in climate model simulations varies. While models indicate that cool SSTs in the eastern tropical Pacific (La Niña or the cool phase of the PDO) are associated with drought in southwestern North America, they also show that atmospheric internal variability may be a more prominent driver (Coats et al., 2015, 2016; Stevenson et al., 2015; Parsons et al., 2018). CMIP5–PMIP3 last millennium simulations reproduce the observed negative correlation between eastern Australian rainfall and the central equatorial Pacific SSTs with varying skill, and also display periods when the ENSO teleconnection weakens substantially for several decades (Brown et al., 2016a). Differences in simulated internal variability have been found to be responsible for the inter-model spread in predicted shifts in subtropical dry zones for a given shift in the Hadley cell (Seviour et al., 2018). CMIP5 models show that both internal variability and anthropogenic forcings are responsible for the drying over the South Atlantic Convergence Zone region, though with large uncertainties (Zilli and Carvalho, 2021). Moreover, the detection of the anthropogenic forcing on the South Atlantic Convergence Zone is strongly dependent on the characterization of model internal variability (Talento and Barreiro, 2012).

Beyond the tropics, North Pacific decadal variability (Annex IV.2.6, 2.4.5, 3.7.6) exerts a strong modulation of extratropical ENSO teleconnections, but also influences low-frequency variability of the Walker circulation, which is underestimated by most CMIP5 models (England et al., 2014). Atlantic Multidecadal Variability (Annex IV.2.7, 2.4.6, 3.7.7) teleconnections show a high model spread among CMIP5 models, both in terms of persistence and spatial coherence (Qasmi et al., 2017), which has potential consequences for the water cycle variability simulated over Europe. For example, internal variability will continue to play an important role in the variability of river flows over France in coming decades *(medium confidence)* (Giuntoli et al., 2013; Boé and Habets, 2014; Bonnet et al., 2017).

Ensembles of atmosphere-only simulations driven by observed or reconstructed SST are useful for evaluating the ability of models to capture the circulation and/or precipitation variability observed over the historical period (Zhou et al., 2016; Deng et al., 2018; Douville et al., 2019). However, limitations of such AGCM-based attribution methods, i.e. related to the lack of air–sea interactions in the response, may lead to erroneous attribution conclusions in some regions for local circulation and mean and extreme precipitation (Dong et al., 2017). Other methods to measure the portion of precipitation variability include the partitioning into dynamical versus thermodynamical components (Saffioti et al., 2016; Fereday et al., 2018; Lehner et al., 2018), the analysis of variance (Dong et al., 2018a) and direct characterization of stochastic weather-noise (Short Gianotti et al., 2014).

Single model initial conditions large ensembles (SMILEs) are a powerful tool for estimating the magnitude of internal variability in historical and future climates (Section 1.4.4). Using SMILEs, it has been shown, for example, that internal NAO variability imparts substantial uncertainty to future changes in European precipitation (Deser et al., 2017; Figure 8.24). For the South Asian summer monsoon, internal variability can overshadow the forced monsoon rainfall trend, thereby increasing near-term projection uncertainties (Huang et al., 2020b). Specific regional applications of the use of large ensembles are further assessed in Sections 10.3.4.3 and 10.3.4.4.

Since AR5, SMILEs have helped quantify the time of emergence of climate change signals (see Sections 1.4.2.2 and 10.4.3). Results from SMILEs indicate that by 2000–2009 (compared to 1950–1999), simulated anthropogenic shifts in mean annual precipitation already emerged over 36-41% of the globe including high latitudes (Frankcombe et al., 2018; Kumar and Ganguly, 2018), the eastern subtropical oceans, and the tropics (Zhang and Delworth, 2018). By 2050 (2100), more than 60% (85%) of the globe is projected to show detectable anthropogenic shifts in mean annual precipitation (Zhang and Delworth, 2018). Other SMILE results for the 1950-2100 period (Kay et al., 2015; Sigmond and Fyfe, 2016) indicate that internal variability can obscure the detection of the anthropogenic hydroclimatic signal until the middle to late 21st century in many parts of the world for both mean and extreme precipitation (Martel et al., 2018; Dai and Bloecker, 2019). A common finding is that changes in the characteristics of wet extreme events will emerge earlier than changes in average conditions (Gaetani et al., 2020; Hawkins et al., 2020; Kusunoki et al., 2020). An assessment of the methods used to estimate time of emergence is presented in Chapter 10 (Section 10.3.4.3). For specific regional examples of climate change attribution and emergence of anthropogenic signal, see Section 10.4.2.

[START FIGURE 8.24 HERE]

Figure 8.24: Impact of the North Atlantic Oscillation (NAO) on 2016–2045 climate trends. (a) Regressions of winter sea level pressure (SLP) and precipitation trends upon the normalized leading principal component (PC)of winter SLP trends in the CESM1 Large Ensemble, multiplied by two to correspond to a two standard deviation anomaly of the PC (as internal climate variability component); (b) CESM1 ensemble-mean winter SLP and precipitation trends (as forced climate variability component); (c) b – a (forced minus internal climate variability component); (d) b + a (forced plus internal climate variability component). Precipitation in colour shading (mm/day per 30 years) and SLP in contours (interval = 1 hPa per 30 years with negative values dashed) (Adapted from Deser et al., 2017). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.24 HERE]

In summary, there is *medium confidence* that climate models reproduce the general magnitude and character of internal variability that influences water cycle variables. There is *high confidence* that internal variability will continue to be a major source of uncertainty, at least for near-term water cycle projections at the regional scale. There is *low confidence* in the region-dependent time of emergence of water cycle changes (see also Section 10.4.3), but there is *medium confidence* that changes in wet extreme events will emerge earlier than changes in average conditions.

8.5.2.2 Implications for near-term water cycle projections

Adapting water resource management in the face of climate change will greatly benefit from improved prediction of land surface hydrology at the decadal timescale. Climate predictions (Section 1.4.4) differ from climate projections by constraining the initial state of the slow components of the climate system (i.e. the ocean, the cryosphere and the terrestrial hydrology) as well as volcanic aerosols and ozone depleting substances with observations. Anthropogenic and natural radiative forcing and low-frequency modes of variability (e.g., AMV and PDV, Annex IV.2.7 and IV.2.6) suggest the possible predictability of climate in the first decade or so of the 21st century, in addition to the projected response to the anthropogenic forcing (Sections 4.2.3 and 4.4.1.3).

In AR5, decadal prediction of precipitation over some land areas showed improved skill due to specified radiative forcing, with almost no added value from ocean initialization. Since AR5, more studies have been devoted to understanding the potential or effective water cycle predictability related to ocean multi-decadal variability. Decadal hindcast experiments based on large ensembles highlight increasing skill scores in annual mean precipitation 3-7 year ahead, at least over the Sahel and Europe (Yeager et al., 2018). There is relatively high predictability of the AMV impacts over the Mediterranean basin, central Asia and the Americas (from United States to northern South America) during boreal summer, but in boreal winter the signal-to-noise ratio shows only weak predictability over land (Yamamoto and Palter, 2016; Ruprich-Robert et al., 2017). The link between South Asian summer monsoon changes and the AMOC and the decadal variability in the Pacific Ocean open the possibility of increased predictability for the near future (Kushnir et al., 2017; Huang et al., 2020c; Sandeep et al., 2020).

The additional skill associated with the initialization of the cryosphere and the land surface has received limited attention. However, there is observational evidence that oceanic decadal variations can propagate into the atmosphere and, consequently accumulate into terrestrial land surface reservoirs (e.g., Bonnet et al., 2017) and vegetation (e.g., Zeng et al., 1999). This land surface memory, like in soil moisture (Alessandri and Navarra, 2008; Catalano et al., 2016) or snow (Loranty et al., 2014), may also contribute to the decadal predictability of the terrestrial component of the water cycle, but remains difficult to assess given the limitations of observational records. Vegetation initialization seems to generate as much noise as signal and does not necessarily translate into improved skill in early decadal predictions based on ESMs (Weiss et al., 2014).

Decadal hydrological predictability in an idealized setting has been also investigated through off-line land surface hindcast experiments, driven by observed atmospheric forcing and/or initial conditions, suggesting the potential for skilful predictions for terrestrial water storage, deep soil moisture, and groundwater (Yuan and Zhu, 2018). Yet, a real-world assessment is hampered by the lack of observations and is only feasible when multi-decadal records of satellite estimates of terrestrial water storage, snow mass or soil moisture are available.

In summary, there is *high confidence* that the water cycle changes that have already emerged from internal variability will become more pronounced in near-term (2021-2040) projections. However, there is *low confidence* in decadal predictions of precipitation changes, particularly over most land areas, because internal variability remains difficult to predict and can offset or amplify the forced water cycle response.

8.5.2.3 Volcanic forcing

Volcanic eruptions can affect climate projections in the near term (2021-2040) (Section 4.4.4, CCB4.1). In this chapter, they are of interest because they can trigger a transient departure from the water cycle response to anthropogenic radiative forcing. Major volcanic eruptions temporarily reduce total global and wet tropical region precipitation (*high confidence*) (Iles et al., 2014), can weaken or shift the ITCZ (Iles et al., 2014; Colose et al., 2016; Liu et al., 2016), and reduce summer monsoon rainfall (*medium confidence*) (Pausata et al., 2015b; Zambri and Robock, 2016; Zambri et al., 2017; Zuo et al., 2019; Singh et al., 2020c). Monsoon precipitation in one hemisphere can be enhanced by the remote volcanic forcing occurring in the other hemisphere (*medium confidence*) (Pausata et al., 2015a; Liu et al., 2016; Zuo et al., 2019). Over the Sahel, the sign of hydrological changes depend on the hemisphere where the volcanic eruptions occur (Haywood et al., 2013b). Out of phase changes in the Sahel and the Amazon basin are expected from the effect of volcanic aerosols on tropical Atlantic SST and the ITCZ (Hua et al., 2019). Over the last millennium, uncertainties remain in the symmetry/asymmetry of the monsoon response because it is difficult to estimate the exact latitude and season of past volcanic eruptions further back in time (Colose et al., 2016; Fasullo et al., 2019).

Data for six major eruptions over the last century along with CMIP5 historical experiments indicate that volcanic eruptions cause a detectable decrease in streamflow in northern South America, central Africa, high-latitude Asia and in wet tropical-subtropical regions, and a detectable increase in southwestern North America and southern South America (Iles and Hegerl, 2015). Attempts to include volcanic forcing in future projections show enhanced precipitation variability on annual to decadal timescales with small reductions in Asian monsoon rainfall (Bethke et al., 2017). The occurrence of volcanic eruptions in the coming century, either as single large events or clustered smaller ones, can alter the water cycle (see also CCB4.1), and regional drought events may be enhanced by co-occurring volcanic (Liu et al., 2016; Gao and Gao, 2017; Zambri et al., 2017) and GHG (e.g., Cook et al., 2018) forcing (*low confidence*). Volcanic eruptions may also lead to widespread precipitation anomalies up to several years following an eruption through their potential influence on the El Niño Southern Oscillation (*low confidence*) (Stevenson et al., 2016; Dee et al., 2020; McGregor et al., 2020).

In summary, large volcanic eruptions reduce global mean precipitation, as well as precipitation in tropical wet regions (*high confidence*). There is *low confidence* in specific regional and seasonal responses, primarily due to the limitations of the observational record.

8.5.3 Non-linearities across global warming levels

AR5 concluded that annual and seasonal mean precipitation changes can be estimated by linear pattern scaling techniques (Santer and Wigley, 1990; Arnell and Gosling, 2016; Greve et al., 2018), which represent regional changes in precipitation as a linear function of global mean temperature change. However, there are a number of caveats when pattern scaling is applied to low emission scenarios or to scenarios where localized forcing (e.g., anthropogenic aerosols) are significant and vary in time (Collins et al., 2013). Here the focus is in on non-linear water cycle responses to increasing global warming levels, as estimated for instance from the difference between the first 2°C of global warming, and the next 2°C of warming (Fig 8.25), and their possible underlying mechanisms.

8.5.3.1 Non-linearities in large-scale atmospheric circulation and precipitation

Since AR5, there is further evidence that the pattern scaling technique has limitations (Lopez et al., 2014; Wartenburger et al., 2017; Tachiiri et al., 2019), and that alternative approaches such as multiple regressions using the land-sea warming contrast as an additional predictor offer added value (Joshi et al., 2013). The simplest traditional pattern scaling approach approximates future changes by the product of a time-evolving global surface temperature change and a pattern that varies spatially but is constant across time, scenarios, and models. This technique was shown to be more robust across scenarios rather than across models, with better results for temperature compared with precipitation (Tebaldi and Arblaster, 2014; see also Section 4.2.4). One approach which avoids scaling is to consider a period in a different scenario with the same global surface temperature change (Herger et al., 2015). It is attractive as it provides patterns of any temporal resolution that are consistent across variables. Nonetheless, this technique is still only based on global surface temperature and is not necessarily suitable for precipitation changes projected in stabilized versus transient scenarios (at the same global warming level) given the fast-atmospheric adjustment to GHG radiative forcing (Section 8.2.1, 8.4.1.1).

Even in a theoretical climate system governed by linear processes, pattern-scaling assumptions can fail because the different forcing time response of different parts of the Earth system cause evolving spatial warming patterns (Good et al., 2016a). This occurs primarily because different feedbacks occur at different timescales (Armour et al., 2013; Andrews et al., 2015), which in turn implies that the atmospheric circulation and water cycle is dependent both on the level of warming and the rate of change (McInerney and Moyer, 2012; Ceppi et al., 2018). The usual distinction between the fast adjustment to increased GHG concentrations and the slower response to SST warming (Section 8.2.2.2) may however not be sufficient to explain the time evolution of the hydroclimatic response at the regional scale, especially in subtropical land areas where this response critically depends on shifts in atmospheric circulation associated with distinct "fast" (typically 5-10 years, that is however much slower than the atmospheric adjustment assessed in Section 8.2.1) and slow SST warming patterns (Zappa et al., 2020). The changing balance between the water cycle response to anthropogenic GHG and aerosol forcings is another source of non-linearity across time and global warming levels (Ishizaki et al., 2013; Rowell et al., 2015; Liu et al., 2019c; Wilcox et al., 2020).

Nonlinearities in the climate response are thought to arise from multiple factors. These include state-dependent ice-albedo feedback and its potential influence on Northern Hemisphere storm tracks (Peings and Magnusdottir, 2014; Semenov and Latif, 2015; see also Cross-Chapter Box 10.1 and Section 8.6.1.2); a state-dependent sensitivity of tropical precipitation to increased SST (Schewe and Levermann, 2017; He et al., 2018); a complex response of the Atlantic meridional overturning circulation (AMOC; Section 9.2.4.1 and Section 8.6.1.1) and its model- and magnitude-dependent teleconnections with regional temperature and precipitation (Kageyama et al., 2013; Jackson et al., 2015; Qasmi et al., 2017, 2020); and other atmospheric and terrestrial (Section 8.5.3.2) processes such as cloud and land surface feedbacks (Ceppi and Gregory, 2017; King, 2019). The response of convective precipitation may exhibit nonlinearities because it is itself modulated by both dynamics and atmospheric water content, each responding independently to warming (Chadwick and Good, 2013; Neupane and Cook, 2013).

Based on a simple model, it was also suggested that the Indian summer monsoon may exhibit a moisture-advection feedback which allows multiple stable states as boundary conditions change (Zickfeld et al., 2005). However, limitations of this theory and comprehensive GCMs suggest a near-linear monsoon response to a broad range of radiative forcings (Boos and Storelvmo, 2016). Non-linear precipitation responses to global warming have been reported in the Indo-Pacific, where a linear increase in SSTs can trigger nonlinear changes in precipitation and a shift in the ITCZ depending on the relative amplitudes of uniform and structured SST anomalies (Chung et al., 2014a; Toda and Watanabe, 2018).

Compared to atmospheric circulation and seasonal mean precipitation, extreme precipitation has been found to scale more accurately with local and global mean temperature (Chou et al., 2012; Pendergrass et al., 2015). The projected increase in the magnitude of extreme precipitation is generally proportional to the global warming level, with an increase of around 7% per 1°C warming (Section 11.4.5) although this rate shows seasonal and geographical variations and is slightly less for 5-day than for 1-day precipitation maxima. Projected changes in extreme precipitation are the result of both thermodynamical and more model-dependent and potentially less linear dynamical contributions (Pfahl et al., 2017). Projected changes in precipitation extremes are also potentially sensitive to a nonlinear response of spatial convective organization (Pendergrass et al., 2016), and can exhibit a quadratic rather than linear response to global warming (Pendergrass et al., 2019).

Within CMIP6, the linearity to CO_2 forcing can be assessed through the comparison of the model response to abrupt doubling versus abrupt quadrupling of atmospheric CO_2 (Webb et al., 2017). Preliminary analyses based on CMIP5 models showed that annual precipitation changes following a doubling step change in $CO_2$from pre-industrial levels are not necessarily consistent with the response to the step from doubling to quadrupling despite a similar change in radiative forcings (Good et al., 2016a; Ceppi and Shepherd, 2017). Beyond the visual comparison of the climate response at various global warming levels (e.g., Figure 4.35), the linearity across global warming levels can be assessed by using the highest emission scenario and comparing seasonal mean relative precipitation changes at +2°C versus +4°C above preindustrial (1850-1900) temperatures (Figure 8.25). The results support the previous finding (Good et al., 2016b) that a second 2°C warming does not necessarily lead to the same precipitation anomaly pattern as the first 2°C, especially in the tropics where regional differences can be large but not necessarily consistent among different models. They are also consistent with a recent analysis of CMIP5 models showing that the projected drying in the Mediterranean and in Chile is substantially faster than the increase in GSAT, and therefore does not scale linearly with global warming (Zappa et al., 2020).

In summary, there is *high confidence* that continued global warming will further amplify GHG-induced changes in large-scale atmospheric circulation and precipitation. Nonetheless, there are cases where regional water cycle changes are not linearly related to global warming due to the interaction of multiple forcings, feedbacks and timescales (*medium confidence*, see also Section 4.2.4, Section 7.4.3, Section 8.2.1). Aridity in subtropical regions is highly sensitive to fast shifts in large-scale atmospheric circulation so are particularly susceptible to such non-linearities.

[START FIGURE 8.25 HERE]

Figure 8.25: Effect of first versus second 2°C of global warming relative to the 1850-1900 base period on seasonal mean precipitation (mm/day). CMIP6 multi-model ensemble mean DJF (left panels) and JJA (right panels) precipitation difference for a,b) SSP5-8.5 at +2°C; c,d) SSP5-8.5 at +4°C minus SSP5-8.5 at +2°C (second 2°C warming); e,f) second minus first 2°C fast warming (c-a and d-b). Only models reaching the +4°C warming levels in SSP5-8.5 are considered. Differences are computed based on 21-yr time windows centered on the first year reaching or exceeding the selected global warming level using a 21-yr running mean global surface atmospheric temperature criterion. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.25 HERE]

8.5.3.2 Non-linearities in land surface processes and feedbacks

Land surface responses and feedbacks represent a potential source of non-linearity for the water cycle response, at least at regional and local scales. The forced response of soil moisture and freshwater resources not only depends on precipitation, but also on evaporation (Laîné et al., 2014), snowmelt (Thackeray et al., 2016), and runoff (Zhang et al., 2018e) which are intrinsically non-linear processes depending on soil moisture or temperature thresholds. Bare ground evaporation is, for instance, usually estimated as a non-linear function of surface soil moisture (Jefferson and Maxwell, 2015). Plant transpiration requires more complex formulations with non-linear dependencies on multiple environmental factors including root-zone soil moisture and atmospheric CO_2 concentration (Franks et al., 2017). Globally, land surface evaporation is both energy and soil-moisture limited, but one of these limitations can become dominant depending on regions and seasons. Non-linearities may be particularly strong in transitional regimes where and when soil moisture limitation plays a major role (Berg and Sheffield, 2018b).

Snowmelt is a non-linear process and projected changes in snowfall are also a non-linear combination of changes in total precipitation and in the fraction of solid precipitation. In cold regions, snowfall may first increase because of the increased water capacity of a warmer atmosphere and then decrease because snow falls as rain in an even warmer atmosphere. Such non-linearities can contribute to elevation, latitudinal and seasonal contrasts in the observed and projected retreat of the Northern Hemisphere snow cover (Shi and Wang, 2015; Thackeray et al., 2016). Mountain glaciers also represent source of nonlinear runoff responses since the annual runoff can first increase due to additional melting and then decrease as the glaciers shrink (Kraaijenbrink et al., 2017; Shannon et al., 2019). Section 9.5.1.3 concludes with *high confidence* that the average annual runoff from glaciers will generally reach a peak at the latest by the end of the 21st century, and decline thereafter. This peak may have already occurred for small catchments with little ice cover, but tends to occur later in basins with large glaciers. Permafrost thawing is another mechanism which can trigger a non-linear hydrological response in the high-latitudes of the Northern Hemisphere (Walvoord and Kurylyk, 2016), whose magnitude and potential abruptness is assessed in Section 5.4.3.3.

Land surface runoff and groundwater recharge are highly nonlinear process, depending for instance on rainfall intensity, soil infiltration capacity, vertical profile of soil moisture and water table depth. A non-linear relationship between rainfall and groundwater recharge was observed in the tropics where intense seasonal rainfalls associated with internal climate variability contribute disproportionately to recharge (Taylor et al., 2013b; Cuthbert et al., 2019a). Groundwater fluxes in arid regions are generally less responsive to climate variability than in humid regions, which can temporarily buffer climate change impacts on water resources or lead to a long, initially hidden, hydrological responses to global warming (Cuthbert et al., 2019a). Hydrological model simulations driven by individual and combined forcing show that decreased precipitation can cause larger deficits in soil moisture, streamflow and water table depth than other forcings, but also that these factors are not linearly cumulative when applied in combination (Hein et al., 2018). Surface runoff was found to scale only approximately with global warming (Tanaka and Takahashi, 2017). Significant non-linearities were found in the projected annual mean runoff response to global warming in CMIP5 projections, which could not be entirely explained by precipitation changes (Zhang et al., 2018). Similar nonlinear behaviours are found in CMIP6 models over the Amazon, Yangtze, Niger, Euphrates and Mississippi river basins (Figure 8.26), highlighting the need to reassess the assumption of linearity when estimating regional water cycle changes.

Beyond changes in land surface water fluxes, nonlinearities in the response of soil moisture and freshwater reservoirs have not been well documented in global climate projections but deserve further attention given the complex interactions between the water, energy and carbon cycles (Berg and Sheffield, 2018a), the growing direct human influence on rivers and groundwater (Abbott et al., 2019), and a possible offset between the linear components of changes in precipitation and evapotranspiration. Significant nonlinearities were found in water scarcity projections, as seen by the stronger sensitivity to the first 2°C increase in global warming (Gosling and Arnell, 2016).

In summary, there is both numerical and process-based evidence that terrestrial water cycle changes can be nonlinear at the regional scale (*high confidence*). Nonlinear regional responses of runoff, groundwater recharge and water scarcity have been documented based on both CMIP5 and CMIP6 models, and highlight the limitations of simple pattern-scaling techniques (*medium confidence*). Water resources fed by melting glaciers are particularly exposed to such nonlinearities (*high confidence*).

[START FIGURE 8.26 HERE]

Figure 8.26: Rate of change in basin-scale annual mean runoff with increasing global warming levels. Relative changes (%) in basin-averaged annual mean runoff estimated as multi-model ensemble median from a variable subset of CMIP6 models for each SSP over six major river basins: a) Mississipi, b) Danube, c) Lena, d) Amazon, e) Euphrates, f) Yangtze, g) Niger, h) Indus, i) Murray. The basin averages have been estimated after a first-order conservative remapping of the model outputs on the 0.5° by 0.5° river network of (Decharme et al., 2019). The shaded area indicates the 5-95% confidence interval of the ensemble values across all SSPs. Note that the y-axis range differs across basins and is particularly large for Niger and Murray (panels g and i). The number of models considered is specified for each scenario in the legend located inside panel b. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.26 HERE]

8.6 What is the potential for abrupt change?

In this report, *abrupt change* is defined as a regional-to-global scale change in the climate system that occurs faster than the typical rate of changes in its history, implying non-linearity in the climate response (see Annex VII: Glossary). Often, abrupt change arises from positive feedbacks in the climate system that cause the current state to become unstable, and cross a 'tipping point' (Lenton et al., 2008); i.e., a rapid shift from one climate state to another. The water cycle has several attributes with potential to produce abrupt change. Non-linear interactions between the ocean, atmosphere, and land surface can result in rapid shifts between wet and dry states (Sections 8.6.1 and 8.6.2). Cessation of solar radiation modification could also result in abrupt changes in the water cycle (Section 8.6.3). This section reviews these types of abrupt shifts and assesses the likelihood that they will occur by 2100.

8.6.1 Abrupt water cycle responses to a collapse of Atlantic Meridional Overturning Circulation

Multiple lines of evidence, including both paleoclimate reconstructions and simulations, suggest that a severe weakening or collapse of Atlantic Meridional Overturning Circulation (AMOC, see Glossary) causes abrupt and profound changes in the global hydrological cycle (Chiang and Bitz, 2005; Broccoli et al., 2006; Chiang and Friedman, 2012; Jackson et al., 2015; Renssen et al., 2018). Deep water formation in the North Atlantic is dependent on a delicate balance of heat and salt fluxes (Buckley and Marshall, 2016); disruption in either of these due to melting ice sheets, a change in precipitation and evaporation, or ocean circulation can force AMOC to cross a tipping point (Drijfhout et al., 2015) (SROCC). During the last deglacial transition, one such slowdown in AMOC—during the Younger Dryas event (12,800–11,700 years ago)—caused worldwide changes in precipitation patterns. These included a southward migration of the tropical ITCZ (Peterson et al., 2000; McGee et al., 2014; Schneider et al., 2014; Mohtadi et al., 2016; Reimi and Marcantonio, 2016; Wang et al., 2017c) and systematic weakening of the African and Asian monsoons (Tierney and deMenocal, 2013; Otto-Bliesner et al., 2014; Cheng et al., 2016; Grandey et al., 2016; Wurtzel et al., 2018). Conversely, the Southern Hemisphere monsoon systems intensified (Cruz et al., 2005; Ayliffe et al., 2013; Stríkis et al., 2015, 2018a; Campos et al., 2019). Drying occurred in Mesoamerica (Lachniet et al., 2013) while the North American monsoon system was largely unaffected (Bhattacharya et al., 2018). The mid-latitude region in North America was wetter (Polyak et al., 2004; Grimm et al., 2006; Wagner et al., 2010; Voelker et al., 2015), while Europe was drier (Genty et al., 2006; Rach et al., 2017; Naughton et al., 2019). A transient coupled climate model simulation was able to reproduce the large-scale precipitation response to such an event (Liu et al., 2009) (Figure 8.27a).

These patterns of past hydroclimatic change are relevant for future projections because it is *very likely* that AMOC will weaken by 2100 in response to increased greenhouse gas emissions (Weaver et al., 2012; Drijfhout et al., 2015; Bakker et al., 2016; Reintges et al., 2017) (See also Section 9.2.3.1). Furthermore, there is *medium confidence* that the decline in AMOC will not involve an abrupt collapse before 2100

(Section 9.2.3.1). The response of precipitation to hypothetical AMOC collapse under elevated greenhouse gases bears resemblance to the paleoclimate response during the Younger Dryas event, with some important differences due to effects of increased CO_2 on global precipitation patterns (Figure 8.27b). As with the paleoclimate events, AMOC collapse results in a southward shift in the ITCZ that is most pronounced in the tropical Atlantic. This could cause drying in the Sahel region (Defrance et al., 2017) as well as Mesoamerica and northern Amazonia (Parsons et al., 2014; Chen et al., 2018c). AMOC collapse also causes the Asian monsoon systems to weaken (Liu et al., 2017b) (Figure 8.27b) counteracting the strengthening expected in response to elevated greenhouse gases (see Section 8.4.2). Europe is projected to experience moderate drying in response to AMOC collapse (Jackson et al., 2015).

[START FIGURE 8.27 HERE]

Figure 8.27: (a) Model simulation of precipitation response to the Younger Dryas event relative to the preceding warm Bølling-Allerød period (base colours, calculated as the difference between 12,600–11,700 yr BP and 14,500–12,900 BP from the TraCE paleoclimate simulation of Liu et al., (2009)), with paleoclimate proxy evidence superimposed on top (dots). (b) Model simulation of precipitation response to an abrupt collapse in AMOC under a doubling of 1990 CO_2 levels (after Liu et al., (2017)). Regions with rainfall rates below 20 mm/year are masked. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

[END FIGURE 8.27 HERE]

In summary, given that there is *medium confidence* that the decline in AMOC will not involve an abrupt collapse before 2100, there is *low confidence* that an AMOC-driven abrupt change in the water cycle will occur by 2100. However, if AMOC collapse does occur, it is *very likely* that there would be large regional impacts on the water cycle.

8.6.2 Abrupt water cycle responses to changes in the land surface

Changes in the land surface, including vegetation cover and dust emissions, can trigger abrupt changes in the water cycle. Plants regulate the exchange of water and energy between the land surface and the atmosphere (Section 8.2.3.3), such that sudden shifts in plant functions, types, or biomes can trigger feedbacks that have the potential to cause abrupt changes in the regional water cycle. Dust emissions, from either climatic or land use changes, affect the radiation budget and can regionally exacerbate dry extremes. Below, we assess the likelihood of abrupt changes in the water cycle for the well-studied regions of the Amazon and the Sahel, and the potential for dust emissions to amplify drought and aridity.

8.6.2.1 Amazon deforestation and drying

The Amazon forest plays an active role in driving atmospheric moisture transport and generating precipitation in the South American region (Drumond et al., 2014; Poveda et al., 2014; Yin et al., 2014; Staal et al., 2018, 2020; Agudelo et al., 2019; Espinoza et al., 2019) (SRCCL). This close association between the land surface and the water cycle makes the Amazon a potential hotspot for abrupt change (Torres and Marengo, 2014). Both deforestation and drying are projected to increase by 2100, resulting in a worst-case scenario of up to a 50% loss in forest cover by 2050 (Soares-Filho et al., 2006; Boisier et al., 2015; Steege et al., 2015; Gomes et al., 2019). Deforestation in the Amazon also raises the probability of catastrophic fires (Brando et al., 2014). The combination of deforestation, drier conditions, and increased fire can push the rainforest ecosystem past a tipping point, beyond which there is rapid land surface degradation, a sharp reduction in atmospheric moisture recycling, an increase in the fraction of precipitation that runs off, and a further shift towards a drier climate (Staal et al., 2015; Boers et al., 2017; Zemp et al., 2017; Ruiz-Vásquez et al., 2020). A rapid drop in precipitation has a direct impact on river flows, driving basin-scale shifts from a regulated to unregulated state (Salazar et al., 2018). Regional climate modeling experiments confirm that increased deforestation leads to a drier climate, although not all models show a true tipping point, at least under present-day climatic conditions (Lejeune et al., 2015; Spracklen and Garcia-Carreras, 2015).

In AR5, some simulations using a coupled climate-carbon cycle model exhibited an abrupt dieback of the Amazon forest in future climate scenarios (Oyama and Nobre, 2003; Cox et al., 2004; Malhi et al., 2008). However, subsequent work demonstrated that abrupt Amazon dieback does not occur consistently across, or even within, Earth system models (Lambert et al., 2013; Boulton et al., 2017). The occurrence of dieback is highly dependent on both how dry the simulated climate is in the present day (Malhi et al., 2009) as well as the representation of forest structure and competitive dynamics (Levine et al., 2016). Models with a low diversity of plant characteristics and types have a higher tendency for abrupt change (Sakschewski et al., 2016). Abrupt shifts and ecosystem disruptions can occur on the sub-regional level (Pires and Costa, 2013), highlighting the need for higher-resolution modelling studies. Since AR5, CMIP6 projections suggest that a tipping point in the Amazon system may be crossed on a local or regional scale (Staal et al., 2020) but continue to be highly dependent on model biases in precipitation and the simulation of the land surface. Consequently, the timing, and probability, of an abrupt shift remains difficult to ascertain.

In summary, while there is a strong theoretical expectation that Amazon drying and deforestation can cause a rapid change in the regional water cycle, currently there is *limited* model *evidence* to verify this response, hence there is *low confidence* that such a change will occur by 2100.

8.6.2.2 Greening of the Sahara and the Sahel

Greening of the Sahara and Sahel regions in North Africa, in response to an increase in precipitation, has long been considered an amplifying mechanism that can lead to abrupt change. Although the high surface albedo of the desert stabilizes the energy balance of the system (Charney, 1975), greening can induce strong, positive feedbacks between the land surface and precipitation that can shift the region into a "Green Sahara" state. The fact that the transition phase between a Desert Sahara and Green Sahara is not theoretically stable (Brovkin et al., 1998) creates a tipping point and allows for the possibility of an abrupt shift between dry and wet climate regimes. Paleoclimate reconstructions provide evidence of past "Green Sahara" states (deMenocal and Tierney, 2012), under which rainfall rates increased by an order of magnitude (Tierney et al., 2017), leading to a vegetated landscape (Jolly et al., 1998) with large lake basins (Gasse, 2000; Drake and Bristow, 2006). The underlying driver of the Green Sahara is the periodic increase in summer insolation associated with the orbital precession cycle (Kutzbach, 1981). In this sense, Green Saharas are not direct analogues for a response to anthropogenic greenhouse gas emissions, as these past states were forced by natural, seasonal changes in solar radiation. However, the climate dynamics of Green Saharas (which have global impacts, Pausata et al., (2020)), and the speed of the transitions between Desert Saharas and Green Saharas, are relevant for future projections.

Since AR5, paleoclimatic studies have improved our view of the timing, spatial extent, and speed of transitions associated with the early Holocene (11,000–5,000 yr BP) Green Sahara. Observed transitions into and out of Green Sahara states are always faster than the underlying forcing, in agreement with theoretical considerations (*high confidence*) (Tierney and deMenocal, 2013; Shanahan et al., 2015; Tierney et al., 2017). However, there is *low confidence* in the duration of the transition because sedimentary records cannot typically resolve changes on decadal—multi-decadal timescales (Tierney and deMenocal, 2013). Both paleoclimate data and modelling experiments suggest that the timing and speed of the transition was spatially heterogeneous (*high confidence*), with northern Saharan locations becoming drier thousands of years before more equatorial locations (Shanahan et al., 2015; Tierney et al., 2017; Dallmeyer et al., 2020). These observations are consistent with theoretical studies suggesting that spatial heterogeneity and diversity in ecosystems can mitigate the probability of catastrophic change (Van Nes and Scheffer, 2005; Bathiany et al., 2013). Conversely, low ecosystem diversity can produce local or regional "hot spots" of abrupt change such as those seen in some paleoclimate records (Claussen et al., 2013).

CMIP5 and CMIP6 models**,** some of which include dynamic vegetation schemes, cannot simulate the magnitude, nor the spatial extent, of greening and precipitation change associated with the last Green Sahara under standard mid-Holocene (6 ka) boundary conditions (*high confidence*) (Harrison et al., 2014; Tierney et al., 2017; Brierley et al., 2020) (Figure 3.11). This result remains unchanged since AR4 (Jansen et al., 2007). This may be due to climatological biases in the models (Harrison et al., 2015) or could imply that the strength of the feedbacks between vegetation and the water cycle in the models is too weak (Hopcroft et al., 2017). To date, climate models still only produce the amount and spatial extent of rainfall that is needed to sustain a Green Sahara if they are given prescribed changes in the land surface, such as albedo, soil moisture, vegetation cover and/or dust emissions (Pausata et al., 2016; Skinner and Poulsen, 2016; Tierney et al., 2017).

Some climate model simulations suggest that under future high-emissions scenarios, CO_2 radiative forcing causes rapid greening in the Sahel and Sahara regions via precipitation change (Claussen et al., 2003; Drijfhout et al., 2015). For example, in the BNU-ESM RCP8.5 simulation, the change is abrupt with the percentage of bare soil dropping from 45% to 15%, and percentage of tree cover rising from 50% to 75%, within 10 years (2050-2060) (Drijfhout et al., 2015). However, other modelling results suggest that this may be a short-lived response to CO_2 fertilization (Bathiany et al., 2014).

In summary, given outstanding uncertainties in how well the current generation of climate models capture land-surface feedbacks in the Sahel and Sahara, there is *low confidence* that an abrupt change to a greener state will occur in these regions before 2100 or 2300.

8.6.2.3 Amplification of drought by dust

Mineral dust aerosols in the climate system originate from both semi-permanent and transient sources (Prospero et al., 2002; Ginoux et al., 2012). The former are typically arid regions where significant alluvial sediments have accumulated over time, while the latter are often associated with natural (e.g., droughts, wildfires) and anthropogenic (e.g. land use change, desertification) disturbances. Modern-day dust emissions are dominated by natural sources (Ginoux et al., 2012), although human emissions may contribute 10–60% of the global atmospheric dust load (Webb and Pierre, 2018). Paleo-dust records suggest that human factors (land use change and landscape disturbance) may have doubled global dust emissions between1750 and the last quarter of the 20th century (Hooper and Marx, 2018) (Section 2.2.6).

Dust aerosols influence the climate system and hydrologic cycle through both direct impacts on radiation (absorbing and scattering longwave and shortwave) and via indirect effects on cloud and precipitation processes (Choobari et al., 2014; Kok et al., 2018; Schepanski, 2018) (Box 8.1). The capacity of dust aerosols to suppress precipitation by reducing humidity and energy availability, and increasing stability in the atmosphere (Cook et al., 2013; Huang et al., 2014) can drive positive feedbacks (see also Section 6.3.6). Thus there is strong potential for dust to contribute to abrupt changes in the water cycle, especially in semi-arid regions where wind erosion is highly sensitive to vegetation cover and drought variability (Yu et al., 2015). One such event occurred over the Central United States during the 1930s: the Dust Bowl drought, an iconic event characterized by widespread land degradation and historically unprecedented levels of dust storm activity (Hansen and Libecap, 2004; Lee and Gill, 2015). While initialized by warm sea-surface temperatures in the North Atlantic, modeling work indicates that land cover changes and resulting dust emissions contributed to the severity and spatial extent of the drought by further suppressing precipitation (Cook et al., 2009; Hu et al., 2018; Cowan et al., 2020). There is also increasing evidence that dust aerosol feedbacks are necessary to explain the magnitude of rainfall increase during the mid-Holocene Green Sahara (Pausata et al., 2016; Tierney et al., 2017).

The importance of dust aerosol feedbacks in future abrupt climate events, like droughts or rapid aridification, is unclear. In part, this is because the response of dust aerosol emissions and loading levels in the atmosphere to climate change is highly uncertain (Tegen and Schepanski, 2018; Webb and Pierre, 2018). This difficulty in predicting future dust responses is rooted in the fact that emissions depend on both changes to the land surface (e.g., land use/land cover change, aridification, ecological responses to climate change) and the state of the atmosphere (Tegen and Schepanski, 2018). While there is some evidence that global dust aerosol concentrations in the future will increase (Allen et al., 2016; Tegen and Schepanski, 2018), it is highly dependent on changes in precipitation patterns and atmospheric circulation (see the SRCCL, Section 2.4.1), and it is not clear what the radiative impact will be (Allen et al., 2016; Kok et al., 2018).

In summary, due to *limited evidence*, there is *low confidence* regarding the role of dust in abrupt climate change events over the next century.

8.6.3 Abrupt water cycle responses to initiation or termination of solar radiation modification

Solar radiation modification (SRM) techniques seek to reduce the impacts of climate change by modifying the Earth's radiation budget, either by reflecting incoming solar radiation or increasing the amount of heat lost to space. Note that, following SR1.5, the definition of SRM in this report refers to changes in both solar and longwave radiation (Section 4.6.3.3 and Annex VII: Glossary). A variety of methods have been proposed, including injection of aerosols or their precursors into the stratosphere, cloud brightening, and cirrus cloud thinning (Table 4.8). Since SRM alters the planetary energy balance, changes in the hydrological cycle are theoretically expected (Section 8.2). These changes can be abrupt if the initial magnitude of SRM is large, rather than increased gradually. Since AR5, a diversity of SRM techniques have been tested using climate model simulations, with an increasing focus on consequences for regional water availability. Techniques targeting shortwave radiation (sulfate injection, surface albedo modification, cloud brightening) are *likely* to reduce global mean precipitation relative to future CO_2 emissions scenarios (Bala et al., 2008; Jones et al., 2013a; Tilmes et al., 2013; Ferraro et al., 2014; Crook et al., 2015). In contrast, cirrus cloud thinning, a longwave technique, results in increased global precipitation as it causes enhanced radiative cooling in the troposphere (*medium confidence*) (Crook et al., 2015; Kristjánsson et al., 2015; Jackson et al., 2016).

The magnitude of hydrological disruption for both the initiation and termination of SRM depends on the method used, as well as the strength and duration of its implementation (Ekholm and Korhonen, 2016; Irvine et al., 2019). Under abrupt SRM implementation, hydrological shifts are rapid, occurring within the first decade (Crook et al., 2015). Artificial enhancement of albedo in Northern Hemisphere desert regions causes a southward shift in the Hadley Cell and ITCZ, and extreme drying in the northern tropics (Crook et al., 2015). Uniform or tropical stratospheric sulfate injection weakens the African and Asian summer monsoons and causes drying in the Amazon (Robock et al., 2008; Crook et al., 2015; Dagon and Schrag, 2016). Changes in evapotranspiration can produce large deficits or surpluses in soil moisture and runoff in different regions and seasons (Dagon and Schrag, 2016).

Rapid changes (years to decades) in the hydrological cycle are also expected if SRM is terminated abruptly, either purposefully or because of technical failure or political disagreement. We reiterate the AR5 conclusion that, if SRM "were terminated for any reason, there is *high confidence* that surface temperatures would increase rapidly (within a decade or two) to values consistent with the GHG forcing." The additional global warming caused by SRM termination may result in a rapid increase in global mean precipitation (*medium confidence*) (Jones et al., 2013). Heterogenous regional and seasonal changes are also expected, but are model-dependent (Jones et al., 2013a). As with SRM initiation, the impact of SRM termination is expected to be dependent on the technique deployed.

In summary, it is *very likely* that abrupt water cycle changes will occur if SRM is abruptly initiated or halted, especially in tropical regions. Further assessment of the potential side-effects of SRM is found in Section 4.6.3.3.

8.7 Final remarks

Despite the advances presented in this chapter, there are still many opportunities to improve the understanding and quantification of human influences on past, present and future water cycle changes:

- Extension and development of pre-instrumental data and paleoclimate records, particularly from the

Southern Hemisphere, will improve estimates of the range of natural climate variability and extremes, and our knowledge of how the water cycle responded in past high CO_2 climates.

- Development of longer observational time series that will improve our understanding of physical processes and the analysis and simulation of natural modes of weather and climate variability.

- The use of large model ensembles will help better understand the interactions between climate change and internal variability and in the detection and attribution of observed water cycle changes.

- The simulation of precipitation, latent heating and radiative effects of deep convective clouds would greatly benefit from a better representation of their interactions with aerosols.

- An improvement of the GCM-simulated precipitation, latent heating and radiative effects of deep convective clouds would benefit from an improved representation of their interactions with aerosols.

- Further research on land surface processes, including groundwater recharge, the role of plant physiological changes, land use change, dams and irrigation, will improve future projections of key aspects of the terrestrial water cycle such as aridity and drought.

- Ongoing efforts to develop higher-resolution, 'convection permitting', regional or global climate models will lead to an improved simulation of clouds and precipitation, their coupling with boundary layer and surface processes, their diurnal cycle and high-frequency variability, and their response to climate change, including extreme precipitation events.

- Further analysis of past and current climate variability alongside future climate change projections will provide physically understood constraints for improving the accuracy of regional water cycle simulations, adding value to the results obtained from global climate models.

- Increased understanding of internal variability and interactions with human-induced change will improve efforts to attribute changes in the water cycle and to understand and anticipate future non-linear change.

Frequently Asked Questions

FAQ 8.1: How does land use change alter the water cycle?

The ways in which humans use and change land cover, for example, by converting fields to urban areas or clearing forests, can affect every aspect of the water cycle. Land-use changes can alter precipitation patterns and how water is absorbed into the ground, flows into streams and rivers, or floods the land surface, as well as how moisture evaporates back into the air. Changes in any of these aspects of the interconnected water cycle can affect the entire cycle and the availability of freshwater resources.

Land use describes the combination of activities and ground cover defining each area of the Earth's continental surface. Altering land use can modify the exchange of water between the atmosphere, soil and sub-surface (FAQ 8.1, Figure 1).

For instance, changes in land cover can affect the ability of soils to soak up surface water (infiltration). When soil loses its capacity to soak up water, precipitation that would normally infiltrate and contribute to groundwater reserves will instead overflow, increasing surface water (runoff) and the likelihood of flooding. For example, changing from vegetation to urban cover can cause water to flow rapidly over buildings, roads and driveways and into drains rather than soaking into the ground. Deforestation over wide areas can also directly reduce soil moisture, evaporation and rainfall locally but can also cause regional temperature changes that affect rainfall patterns.

Extracting water from the ground and river systems for agriculture, industry and drinking water depletes ground water and can increase surface evaporation because water that was previously in the ground is now in direct contact with the atmosphere, being available for evaporation

Changing land use can also alter how wet the soil is, influencing how quickly the ground heats up and cools down and the local water cycle. Drier soils evaporate less water into the air but heat up more in the day. This can lead to warmer, more buoyant plumes of air that can promote cloud development and precipitation if there is enough moisture in the air.

Changes in land use can also modify the amount of tiny aerosol particles in the air. For instance, industrial and domestic activities can contribute to aerosol emissions, as do natural environments such as forests or salt lakes. Aerosols cool down global temperature by blocking out sunlight but can also affect the formation of clouds and therefore the occurrence of precipitation (see FAQ 7.2).

Vegetation plays an important role in soaking up soil moisture and evaporating water into the air (*transpiration*) through tiny holes (*stomata*) that allow the plants to take in carbon dioxide. Some plants are better at retaining water than others, so changes in vegetation can affect how much water infiltrates into the ground, flows into streams and rivers, or is evaporated.

More globally, land-use change is currently responsible for about 15% of the emissions of carbon dioxide from human activities, leading to global warming, which in turn affects precipitation, evaporation, and plant transpiration. In addition, higher atmospheric concentrations of carbon dioxide due to human activities can make plants more efficient at retaining water because the stomata do not need to open so widely. Improved land and water management (e.g., reforestation, sustainable irrigation) can also contribute to reducing climate change and adapting to some of its adverse consequences.

In summary, there is abundant evidence that changes in land use and land cover alter the water cycle globally, regionally and locally, by changing precipitation, evaporation, flooding, ground water, and the availability of fresh water for a variety of uses. Since all the components of the water cycle are connected (and linked to the carbon cycle), changes in land use trickle down to many other components of the water cycle and climate system.

[START FAQ 8.1, FIGURE 1 HERE]

FAQ 8.1, Figure 1: Land-use changes and their consequences on the water cycle. As all the components or the water cycle are tightly connected, changes in one aspect of the cycle affects almost all the cycle.

[END FAQ 8.1, FIGURE 1 HERE]

FAQ 8.2: Will floods become more severe or more frequent as a result of climate change?

A warmer climate increases the amount and intensity of rainfall during wet events, and this is expected to amplify the severity of flooding. However, the link between rainfall and flooding is complex, so while the most severe flooding events are expected to worsen, floods could become rarer in some regions.

Floods are a natural and important part of the water cycle but they can also threaten lives and safety, disrupt human activities, and damage infrastructure. Most inland floods occur when rivers overtop their banks (*fluvial* flooding) or when intense rainfall causes water to build up and overflow locally (*pluvial* flooding). Flooding is also caused by coastal inundation by the sea, rapid seasonal melting of snow, and the accumulation of debris, such as vegetation or ice, that stops water from draining away.

Climate change is already altering the location, frequency and severity of flooding. Close to the coasts, rising sea levels increasingly cause more frequent and severe coastal flooding, and the severity of these floods is exacerbated when combined with heavy rainfall. The heavy and sustained rainfall events responsible for most inland flooding are becoming more intense in many areas as the climate warms because air near Earth's surface can carry around 7% more water in its gas phase (vapour) for each 1°C of warming. This extra moisture is drawn into weather systems, fueling heavier rainfall (FAQ 8.2, Figure 1).

A warming climate also affects wind patterns, how storms form and evolve, and the pathway those storms usually travel. Warming also increases condensation rates, which in turn releases extra heat that can energize storm systems and further intensify rainfall. On the other hand, this energy release can also inhibit the uplift required for cloud development, while increases in particle pollution can delay rainfall but invigorate storms. These changes mean that the character of precipitation events (how often, how long-lasting and how heavy they are) will continue to change as the climate warms.

In addition to climate change, the location, frequency and timing of the heaviest rainfall events and worst flooding depend on natural fluctuations in wind patterns that make some regions unusually wet or dry for months, years, or even decades. These natural variations make it difficult to determine whether heavy rainfall events are changing locally as a result of global warming. However, when natural weather patterns bring heavy and prolonged rainfall in a warmer climate, the intensity is increased by the larger amount of moiture in the air.

An increased intensity and frequency of record-breaking daily rainfall has been detected for much of the land surface where good observational records exist, and this can only be explained by human-caused increases in atmospheric greenhouse gas concentrations. Heavy rainfall is also projected to become more intense in the future for most places. So, where unusually wet weather events or seasons occur, the rainfall amounts are expected to be greater in the future, contributing to more severe flooding.

However, heavier rainfall does not always lead to greater flooding. This is because flooding also depends upon the type of river basin, the surface landscape, the extent and duration of the rainfall, and how wet the ground is before the rainfall event (FAQ 8.2, Figure 1) Some regions will experience a drying in the soil as the climate warms, particularly in sub-tropical climates, which could make floods from a rainfall event less probable because the ground can potentially soak up more of the rain. On the other hand, less frequent but more intense downpours can lead to dry, hard ground that is less able to soak up heavy rainfall when it does occur, resulting in more runoff into lakes, rivers and hollows. Earlier spring snowmelt combined with more precipitation falling as rain rather than snow can trigger flood events in cold regions. Reduced winter snow cover can, in contrast, decrease the chance of flooding arising from the combination of rainfall and rapid snowmelt. Rapid melting of glaciers and snow in a warming climate is already increasing river flow in some regions, but as the volumes of ice diminish, flows will peak and then decline in the future. Flooding is also affected by changes in the management of the land and river systems. For example, clearing forests for agriculture or building cities can make rain water flow more rapidly into rivers or low lying areas. On the other hand, increased extraction of water from rivers can reduce water levels and the likelihood of flooding.

A mix of both increases and decreases in flooding have been observed in some regions and these changes have been attributed to multiple causes, including changes in snowmelt, soil moisture and rainfall. Although we know that a warming climate will intensify rainfall events, local and regional trends are expected to vary in both direction and magnitude as global warming results in multiple, and sometimes counteracting, influences. However, even accounting for the many factors that generate flooding, when weather patterns cause flood events in a warmer future, these floods will be more severe.

[START FAQ 8.2, FIGURE 1 HERE]

FAQ 8.2, Figure 1: Schematic illustrating factors important in determining changes in heavy precipitation and flooding.

[END FAQ 8.2, FIGURE 1 HERE]

FAQ 8.3: What causes droughts, and will climate change make them worse?

Droughts usually begin as a deficit of precipitation, but then propagate to other parts of the water cycle (soils, rivers, snow/ice and water reservoirs). They are also influenced by factors like temperature, vegetation and human land and water management. In a warmer world, evaporation increases, which can make even wet regions more susceptible to drought.

A drought is broadly defined as drier than normal conditions; that is, a moisture deficit relative to the average water availability at a given location and season. Since they are locally defined, a drought in a wet place will not have the same amount of water deficit as a drought in a dry region. Droughts are divided into different categories based on where in the water cycle the moisture deficit occurs: meteorological drought (precipitation), hydrological drought (runoff, streamflow, and reservoir storage), and agricultural or ecological drought (plant stress from a combination of evaporation and low soil moisture). Special categories of drought also exist. For example, a snow drought occurs when winter snowpack levels are below average, which can cause abnormally low streamflow in subsequent seasons. And while many drought events develop slowly over months or years, some events, called flash droughts, can intensify over the course of days or weeks. One such event occurred in 2012 in the midwestern region of North America and had a severe impact on agricultural production, with losses exceeding $30 billion US dollars. Droughts typically only become a concern when they adversely affect people (reducing water available for municipal, industrial, agricultural, or navigational needs) and/or ecosystems (adverse effects on natural flora and fauna). When a drought lasts for a very long time (more than two decades) it is sometimes called a megadrought.

Most droughts begin when precipitation is below normal for an extended period of time (meteorological drought). This typically occurs when high pressure in the atmosphere sets up over a region, reducing cloud formation and precipitation over that area and deflecting away storms. The lack of rainfall then propagates across the water cycle to create agricultural drought in soils and hydrological drought in waterways. Other processes act to amplify or alleviate droughts. For example, if temperatures are abnormally high, evaporation increases, drying out soils and streams and stressing plants beyond what would have occurred from the lack of precipitation alone. Vegetation can play a critical role because it modulates many important hydrologic processes (soil water, evapotranspiration, runoff). Human activities can also determine how severe a drought is. For example, irrigating croplands can reduce the socioeconomic impact of a drought; at the same time, depletion of groundwater in aquifers can make a drought worse.

The effect of climate change on drought varies across regions. In the subtropical regions like the Mediterranean, southern Africa, southwestern Australia and southwestern South America, as well as tropical central America, western Africa and the Amazon basin, precipitation is expected to decline as the world warms, increasing the possibility that drought will occur throughout the year (FAQ 8.3, Figure 1). Warming will decrease snowpack, amplifying drought in regions where snowmelt is an important water resource (such as in southwestern South America). Higher temperatures lead to increased evaporation, resulting in soil drying, increased plant stress, and impacts on agriculture, even in regions where large changes in precipitation are not expected (such as central and northern Europe). If emissions of greenhouse gases are not curtailed, about a third of global land areas are projected to suffer from at least moderate drought by 2100. On the other hand, some areas and seasons (such as high-latitude regions in North America and Asia, and the South Asian monsoon region) may experience increases in precipitation as a result of climate change, which will decrease the likelihood of droughts. FAQ 8.3, Figure 1 highlights the regions where climate change is expected to increase the severity of droughts.

[START FAQ 8.3, FIGURE HERE]

FAQ 8.3, Figure 1: Drought is expected to get worse in the regions highlighted in brown as a consequence of climate change. This pattern is similar regardless of the emissions scenario; however, the magnitude of change increases under higher emissions.

[END OF FAQ 8.3, FIGURE]

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Figures

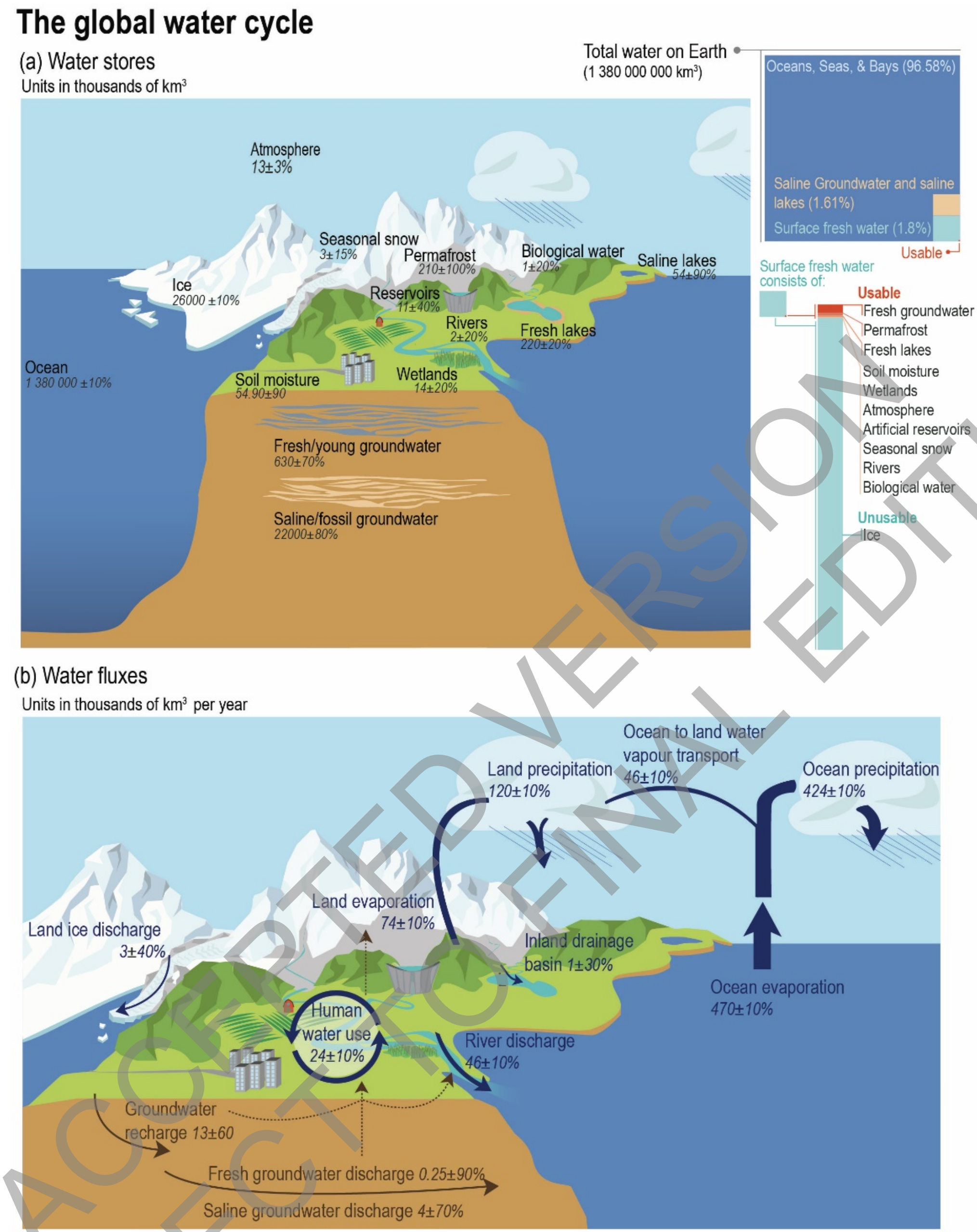


Figure 8.1: Depiction of the water cycle based on previous assessments (Trenberth et al., 2011; Rodell et al., 2015; Abbott et al., 2019) with minor adjustments for groundwater flows (Kwon et al., 2014; Zhou et al., 2019; Luijendijk et al., 2020), seasonal snow (Pulliainen et al., 2020) and ocean precipitation and evaporation (Stephens et al., 2012; Allan et al., 2020; Gutenstein et al., 2020). In the atmosphere, which accounts for only 0.001% of all water on Earth, water primarily occurs as a gas (water vapour), but it is also present as ice and liquid water within clouds. The ocean is the primary water reservoir on Earth, which is comprised mostly of liquid water across much of the globe, but also includes areas covered by ice in polar regions. Liquid freshwater on land forms surface water (lakes, rivers), soil moisture and groundwater stores, together accounting for 1.8% of global water (Stocker et al., 2013). Solid terrestrial water that occurs as ice sheets, glaciers, snow and ice on the surface and permafrost currently represents 2.2% of the planet's water (Stocker et al., 2013). Water that falls as snow in winter provides soil moisture and streamflow after melting, which are essential for human activities and ecosystem functioning.

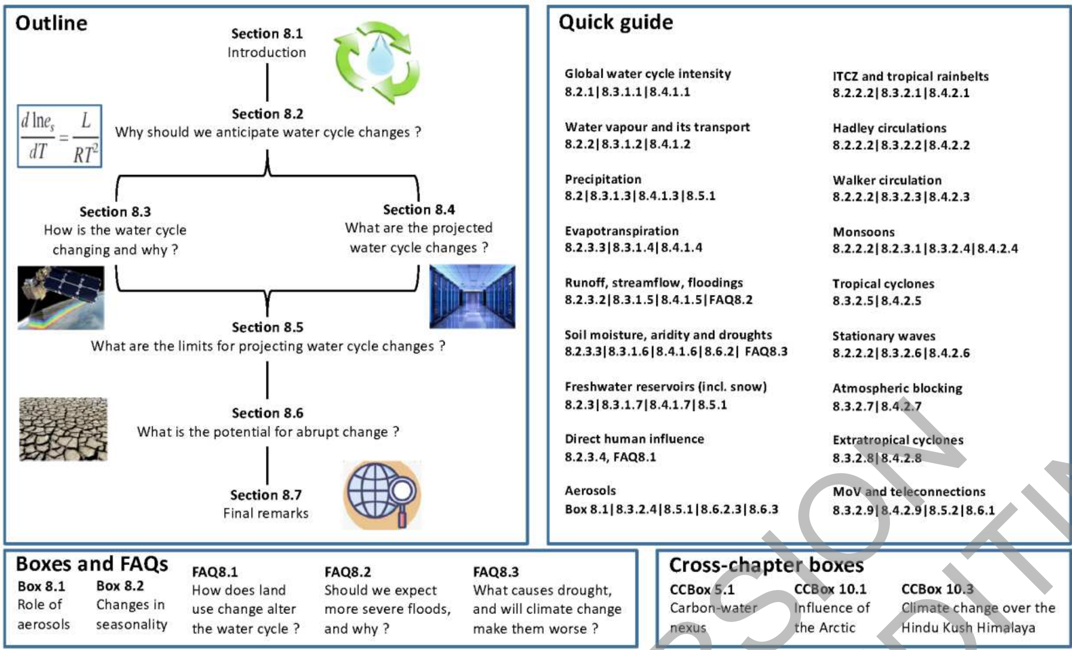


Figure 8.2: Schematic of the chapter structure and quick guide to the chapter content.

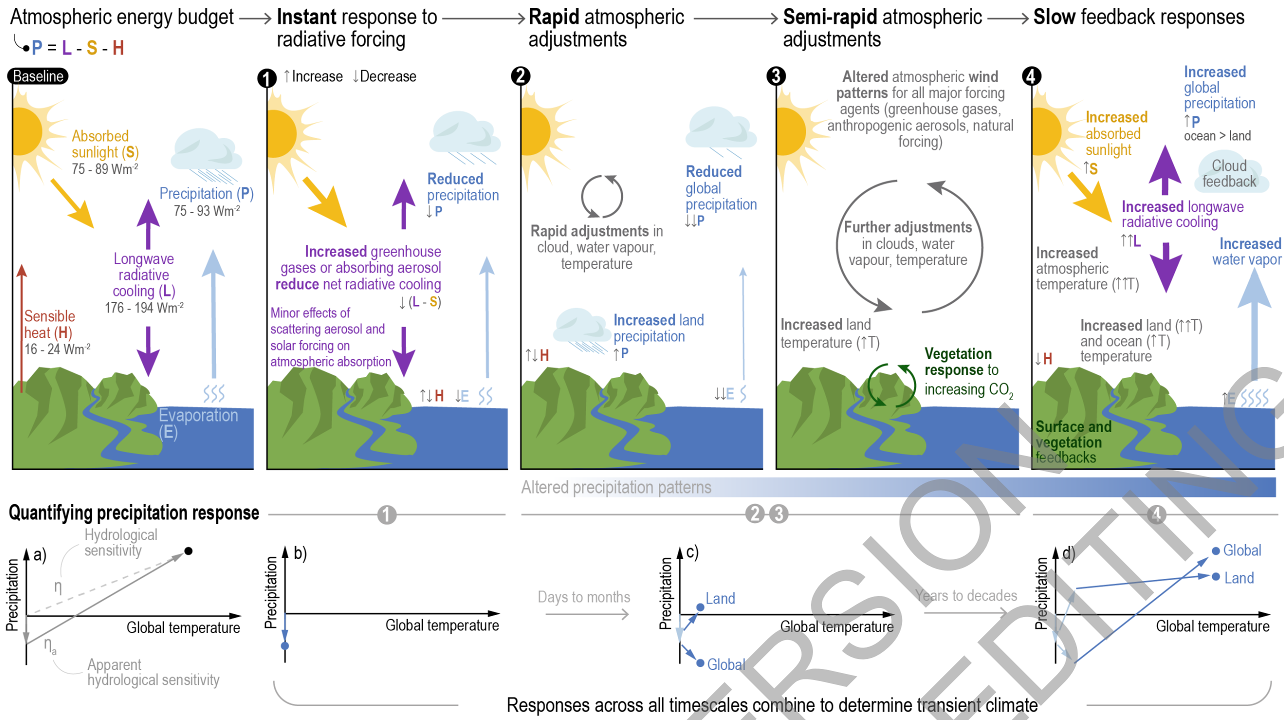


Figure 8.3: Schematic representation of fast and slow responses of the atmospheric energy balance and global precipitation to radiative forcing. The atmospheric energy budget (‘baseline’ panel) responds instantaneously to radiative forcings (1), leading to rapid atmospheric adjustments (2) and slower semi-rapid adjustments involving the land surface and vegetation that further modify atmospheric circulation patterns (3). This slow precipitation response to global mean surface air temperature (4) is quantified as the hydrological sensitivity, η, and the total precipitation response, including initial rapid adjustments, is termed the apparent hydrological sensitivity, η_a (a). The slow precipitation response over land and ocean develops over time (b-d). Large, filled arrows (in panels from 'baseline' to 4) depict fluxes or circulation change while small arrows (1-4) denote increases (↑) or decreases (↓) in variables (P is precipitation; L is atmospheric longwave radiative cooling, S is solar radiation absorption by the atmosphere; H is sensible heat flux; E is surface evaporative heat flux and T is temperature). (Adapted from Allan et al., 2020, Chapter 7 Figure 7.2 and Figure 8.1).

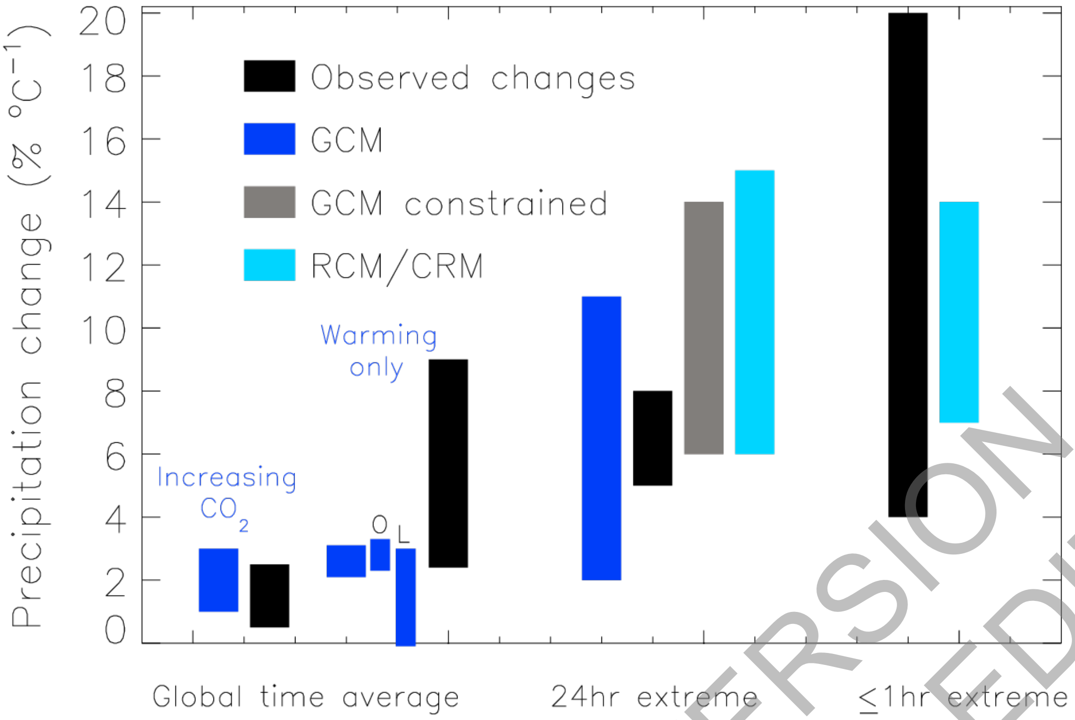


Figure 8.4: Estimate (5-95% range) of the increase in precipitation and its extremes with global mean surface warming. Global time averaged precipitation changes (left) are based on responses to increasing CO_2 (apparent hydrological sensitivity, η_a) and the temperature-dependent component (hydrological sensitivity, η) based on GCM experiments and including the land (L) and ocean (O) components (Fläschner et al., 2016; Richardson et al., 2018; Samset et al., 2018; Pendergrass, 2020; Rehfeld et al., 2020) and observational estimates (GPCP/HadCRUTv4.6) using trends (1988-2014) as a proxy for η_a and interannual variability as a proxy for η with 90% confidence range accounting for statistical uncertainty only (Adler et al., 2017; Allan et al., 2020). For extreme precipitation, assessment is for 24 hour 99.9th percentile or annual maximum extremes from GCMs (Fischer and Knutti, 2015; Pendergrass et al., 2015; Borodina et al., 2017; Pfahl et al., 2017; Sillmann et al., 2017), regional climate models (RCMs) (Bao et al., 2017), an observationally constrained tropical estimate (O'Gorman, 2012) and estimates from observed changes (Westra et al., 2013; Donat et al., 2016; Borodina et al., 2017; Sun et al., 2020; Zeder and Fischer, 2020). For hourly and sub-hourly extremes observed changes (Barbero et al., 2017; Guerreiro et al., 2018) and high resolution models including RCM and cloud resolving models (CRMs) are assessed (Ban et al., 2015; Prein et al., 2017; Haerter and Schlemmer, 2018; Hodnebrog et al., 2019; Lenderink et al., 2019). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

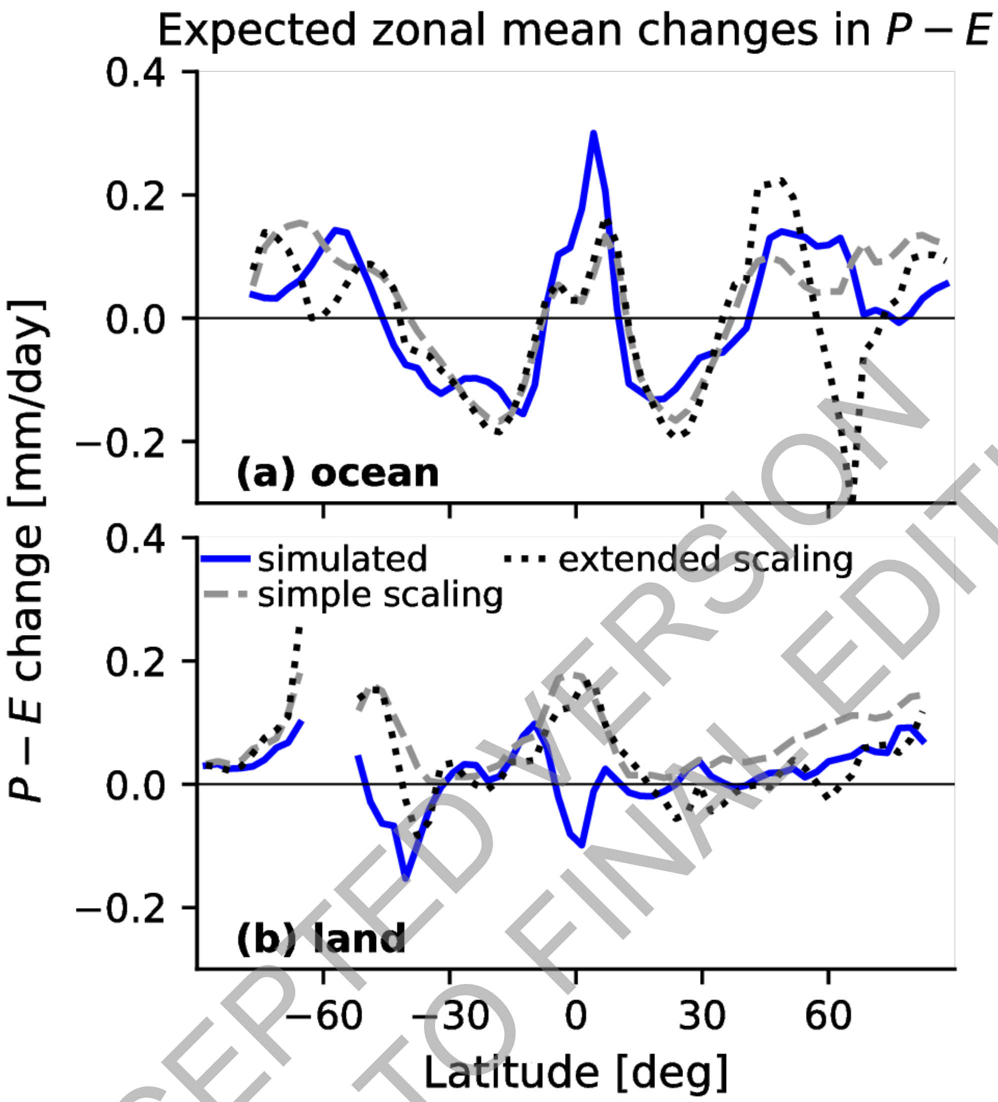


Figure 8.5: Zonally-averaged annual mean changes in precipitation minus evaporation (P-E) over (a) ocean and (b) land between the historical (1995–2014) and SSP2-4.5 (2081–2100) CMIP6 simulations (blue lines, an average of the CanESM5 and MRI-ESM2-0 models). Dashed lines show estimated P-E changes using a simple thermodynamic scaling (Held and Soden, 2006); dotted lines show estimates using an extended scaling (Byrne and O'Gorman, 2016). All curves have been smoothed in latitude using a three grid-point moving-average filter. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

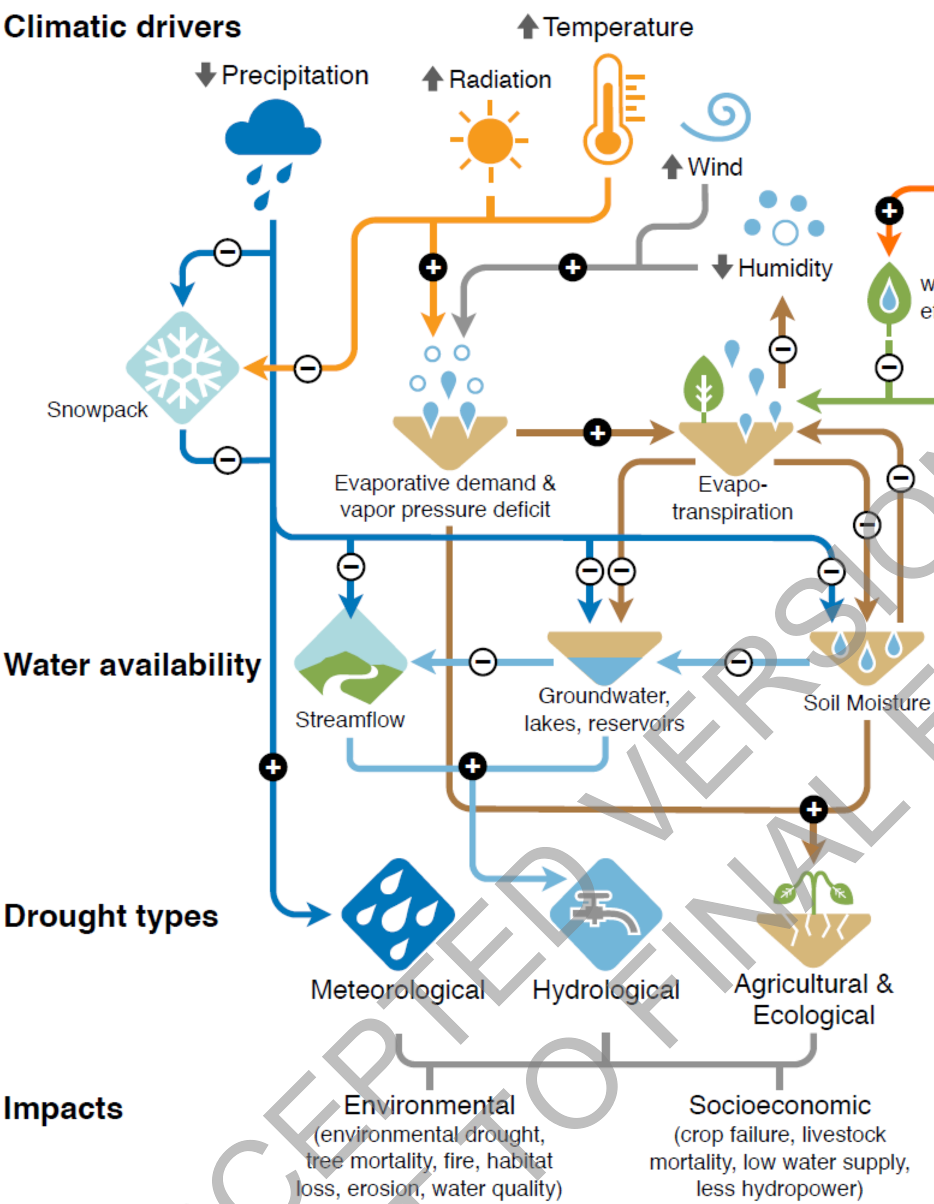
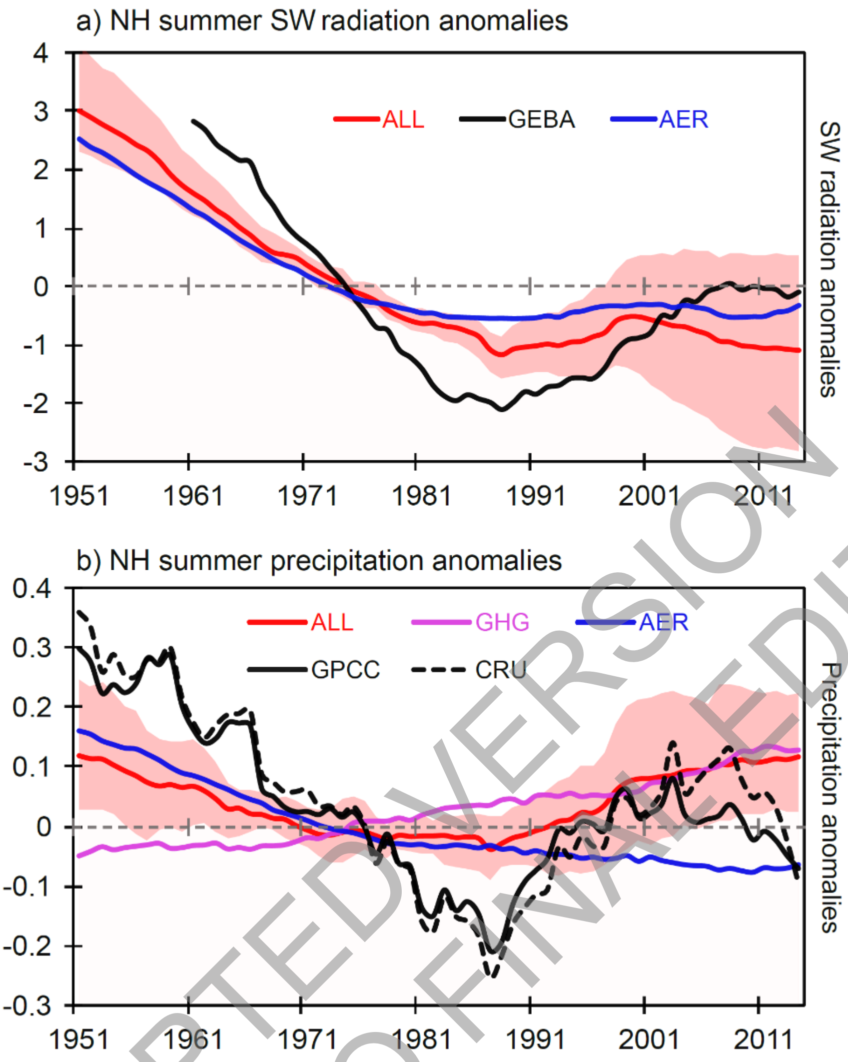
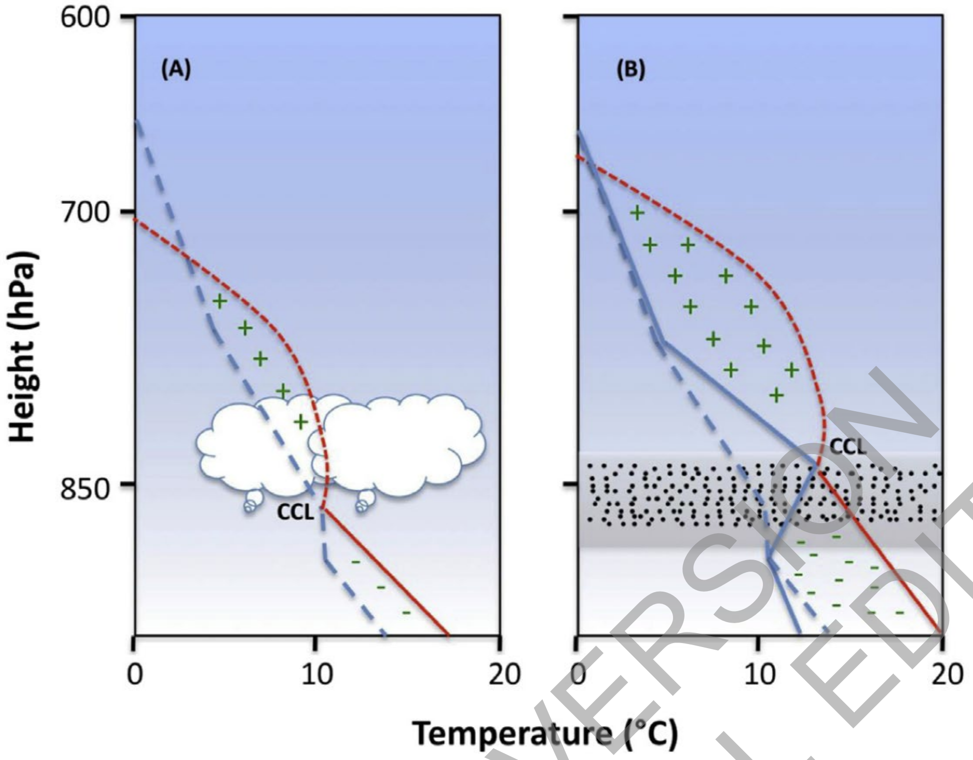


Figure 8.6: Climatic drivers of drought, effects on water availability, and impacts. Plus and minus signs denote the direction of change that drivers have on factors such as snowpack, evapotranspiration, soil moisture, and water storage. The three main types of drought are listed, along with some possible environmental and socioeconomic impacts of drought (bottom).



Box 8.1, Figure 1: Northern hemisphere surface downward radiation anomalies (Wm^{-2} ; a) and precipitation anomalies (mm/day ; b) for 1951–2014 for summer season (May–September) monsoon region (Polson et al. et al., 2014) from CMIP6 DAMIP experiments. Observed solar radiation anomalies are from GEBA global data from 1961-2014 (Wild et al., 2017) and observed precipitation anomalies are from GPCC and CRU. CMIP6 multi-model mean anomalies are from all-forcings (ALL), greenhouse gas forcing (GHG) and anthropogenic aerosol forcing (AER) experiments. Anomalies are with respect to 1961–1990 and smoothed with a 11-year running mean. Red shading shows the ensemble spread of ALL forcing experiment (5%–95% range). Models are masked to the GPCC data set. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).



Box 8.1, Figure 2: Schematic depiction of the atmospheric effects of light absorbing aerosols on convection and cloud formation: (A) without and (B) with the presence of absorbing aerosols in the planetary boundary layer. The dashed and solid blue lines correspond to the vertical temperature profiles in the absence and presence of the absorbing aerosol layer, respectively, and the solid and dashed red lines denote the dry and moist adiabats, respectively. Absorbing aerosols result in an increasing temperature in the atmosphere but a reduced temperature at the surface. The reduced surface temperature and the increased temperature aloft lead to a larger negative energy associated with convective inhibition (-) and a higher convection condensation level (CCL) under the polluted conditions. On the other hand, the absorbing aerosol layer induces a larger convective available potential energy (+) above CCL, facilitating more intensive vertical development of clouds, if lifting is sufficient to overcome the larger convective inhibition. From (Wang et al., 2013).

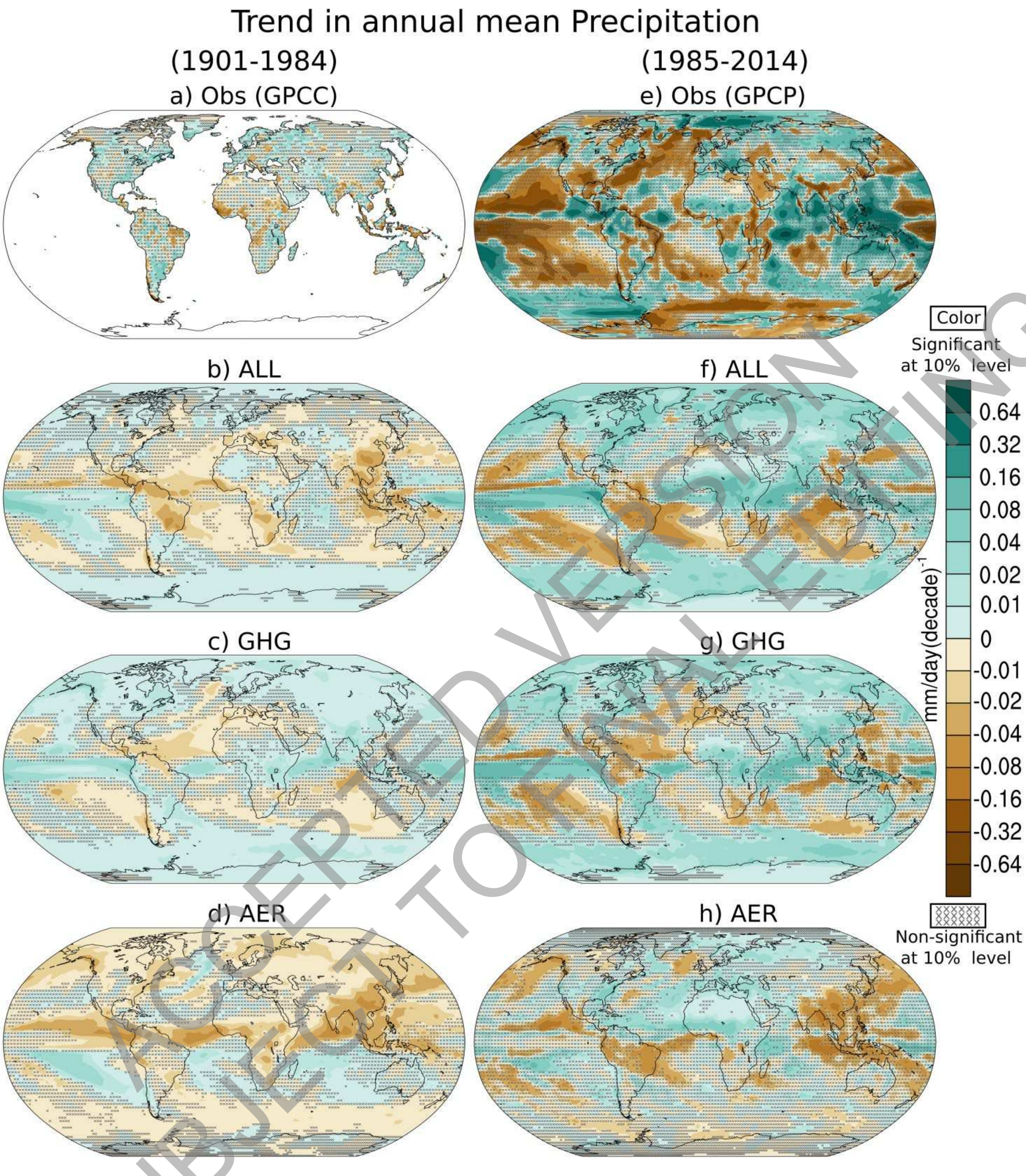


Figure 8.7: Linear trends in annual mean precipitation (mm/day per decade) for 1901-1984 (left) and 1985-2014 (right): (a & e) observational dataset, and the CMIP6 multi model ensemble mean historical simulations driven by, (b & f) all radiative forcings, (c & g) GHG only radiative forcings, (d & h) aerosol only radiative forcings experiment. Colour shades without grey cross correspond to the regions exceeding 10 % significant level. Grey crosses correspond to the regions not reaching the 10% statistical significant level. Nine CMIP6-DAMIP models have been used having at least 3 members. The ensemble mean is weighted per each model on the available and used members. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

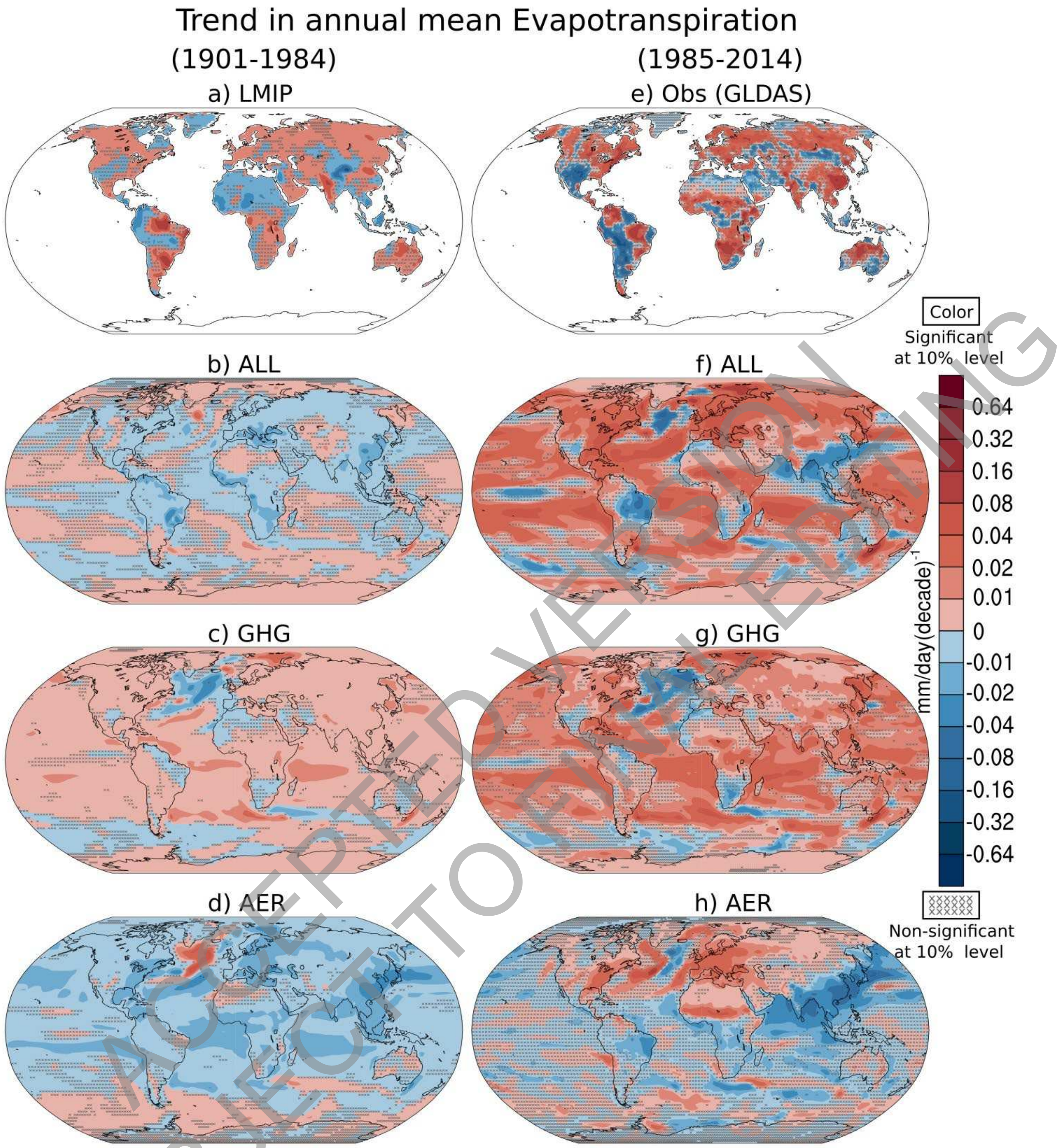


Figure 8.8: Linear trends in annual mean evapotranspiration (mm/day per decade) for 1901-1984 (left) and 1985-2014 (right): (a & e) LMIP and observational dataset, and the CMIP6 multi model ensemble mean historical simulations driven by, (b & f) all radiative forcings, (c & g) GHG only radiative forcings, (d & h) aerosol only radiative forcings experiment. Colour shade without grey cross correspond to the regions exceeding 10 % significant level. Grey crosses correspond to the regions not reaching the 10% statistically significant level. Nine CMIP6-DAMIP models have been used having at least 3 members. The ensemble mean is weighted per each model on the available and used members. GLDAS is not available over the early 20th century so was replaced by a multi-model off-line reconstruction, LMIP, which is consistent with GLDAS over the recent period but may be less reliable over the early 20th century given larger uncertainties in the atmospheric forcings. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

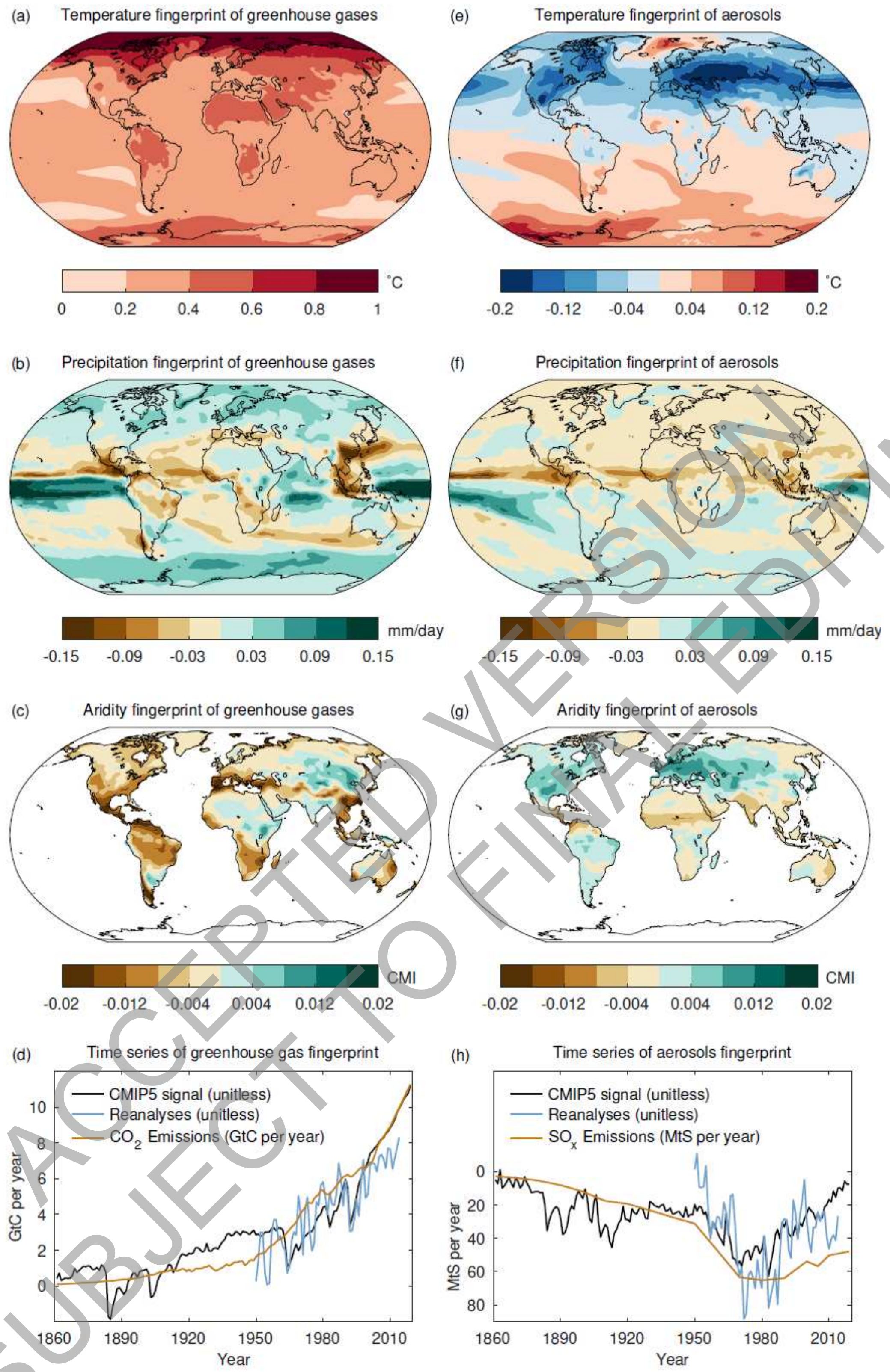


Figure 8.9: Spatial expressions (a-c; e-g) of the leading multivariate fingerprints of temperature (˚C), precipitation (mm/day), and aridity (CMI; the Climate Moisture Index) in CMIP5 historical simulations and the corresponding temporal evolution in both CMIP5 and reanalysis products (d, h). The first leading fingerprint is associated with greenhouse gas forcing (a-d) and the second leading fingerprint is associated with aerosol forcing (e-h). CMI is a dimensionless aridity indicator that combines precipitation and atmospheric evaporative demand. Figure after (Bonfils et al., 2020). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

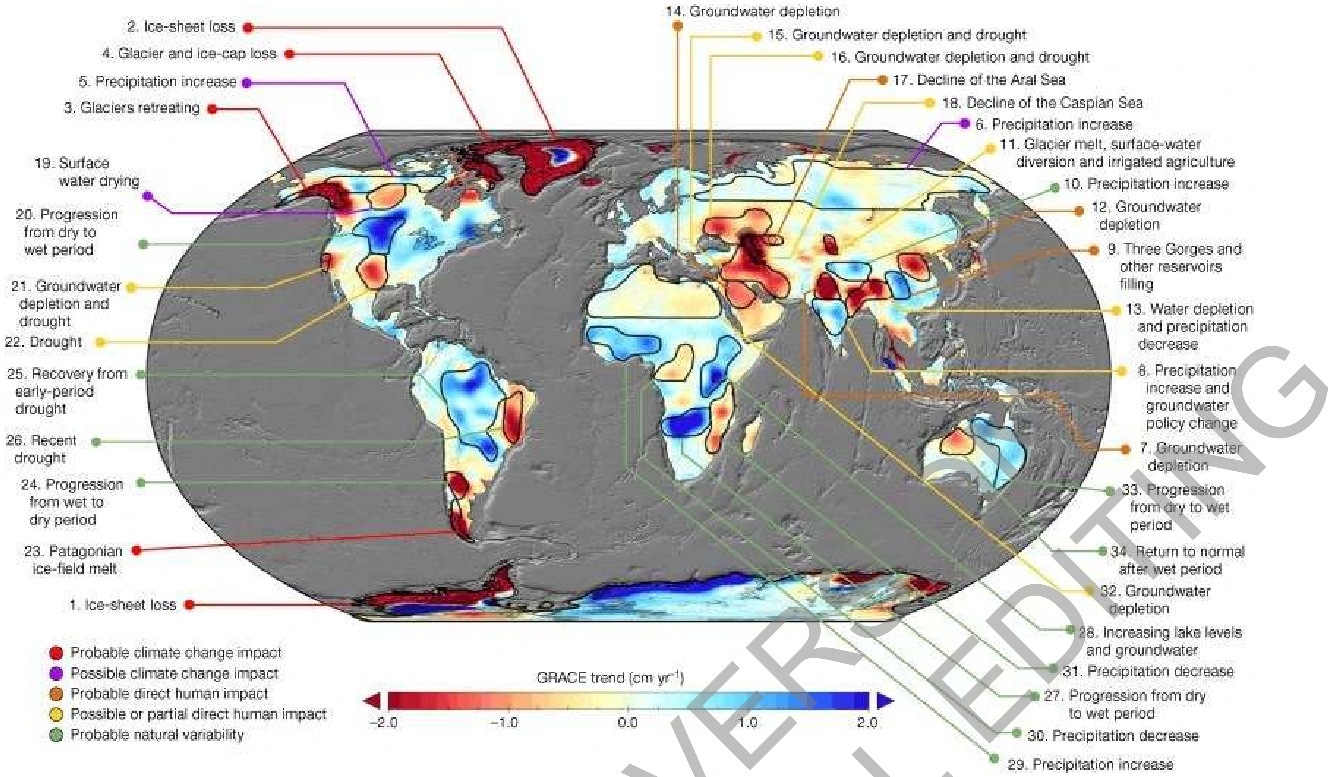


Figure 8.10: Trends in Terrestrial Water Storage (TWS) (in centimetres per year) obtained on the basis of GRACE observations from April 2002 to March 2016. The cause of the trend in each outlined study region is briefly explained and colour-coded by category. The trend map was smoothed with a 150-km-radius Gaussian filter for the purpose of visualization; however, all calculations were performed at the native 3° resolution of the data product. Figure from Rodell et al. (2018). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

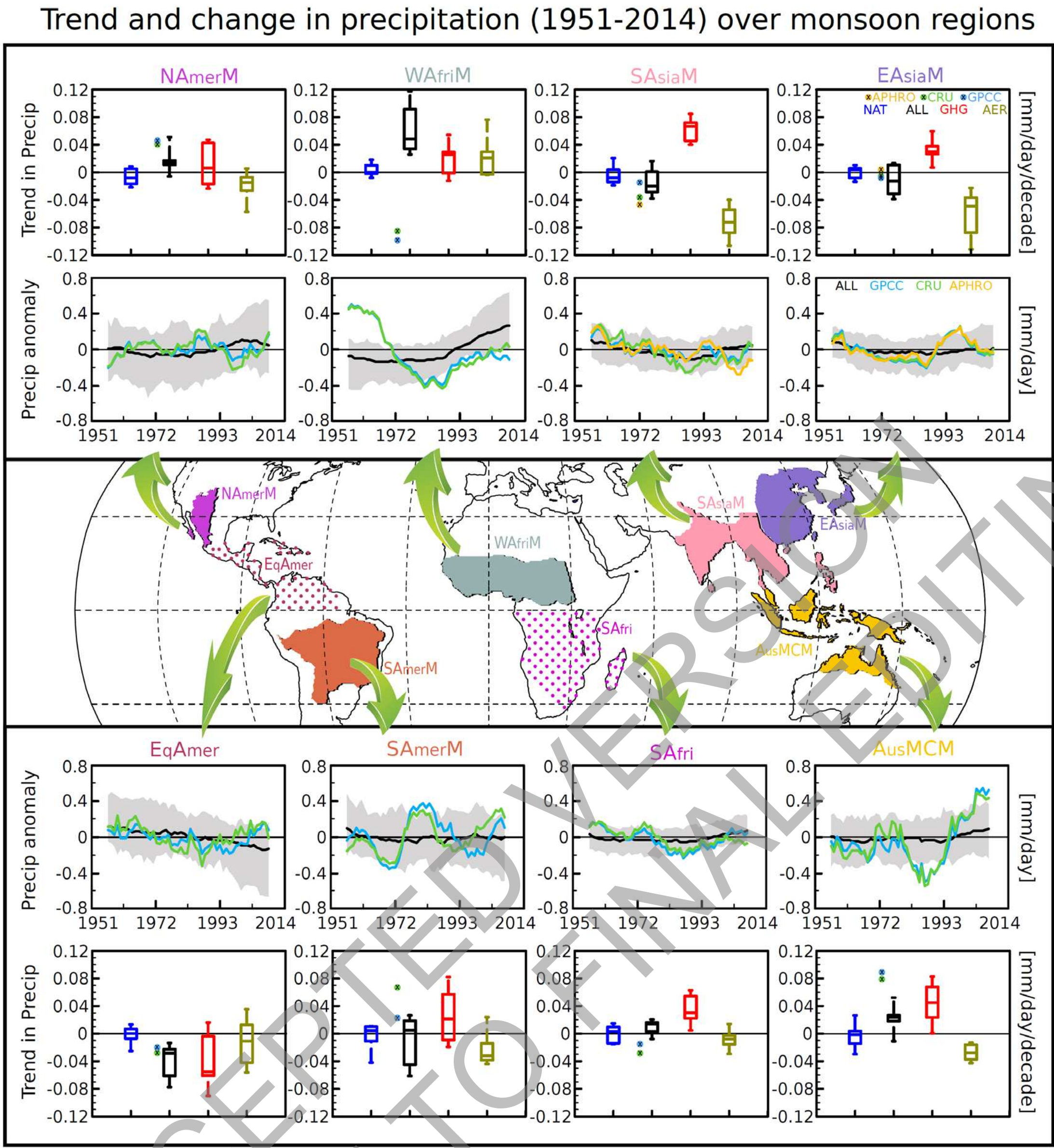


Figure 8.11: Regional monsoon precipitation changes from observations and model attribution. Precipitation changes during 1951-2014 are shown as least-square linear trends in box-whisker plots (first and fourth rows) over the six regional monsoons, i.e., North American monsoon (NAmerM, Jul-Aug-Sep), West African monsoon (WAfriM, Jun-Jul-Aug-Sep), South and Southeast Asian monsoon (SAsiaM, Jun-Jul-Aug-Sep), East Asian monsoon (EAsiaM, Jun-Jul-Aug), South American monsoon (SAmerM, Dec-Jan-Feb), Australian and Maritime Continent monsoon (AusMCM, Dec-Jan-Feb), and over the two land domains (i.e. equatorial America (EqAmer, Jun-Jul-Aug) and South Africa (SAfri, Dec-Jan-Feb), as identified in the map shown in the middle and as described in Annex V. Precipitation changes are computed from observations and from DAMIP CMIP6 experiments over the historical period with all-forcing (ALL), GHG-only forcing (GHG), Aerosol-only (AER) and Natural (NAT) forcings prescribed. Observations are based on the CRU (light green) and GPCC (light blue) datasets and the APHRODITE (light orange) dataset for SAsiaM and EAsiaM. CMIP6 simulations are taken from nine CMIP6 models contributing to DAMIP, with at least 3 members. Ensembles are weight-averaged for the respective model ensemble size. Observed trends are shown as colored cirles and the simulated trends from the CMIP6 multi-model experiments are shown as box-whisker plots. Precipitation anomaly time-series are shown in the second and third row. The thick black line is the multi-model ensemble-mean precipitation anomaly time-series from the ALL experiment and the grey shading shows the spread across the multi-model ensembles. A 11-year running mean has been applied on the precipitation anomaly time-series prior to calculating the multi-model ensemble mean. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

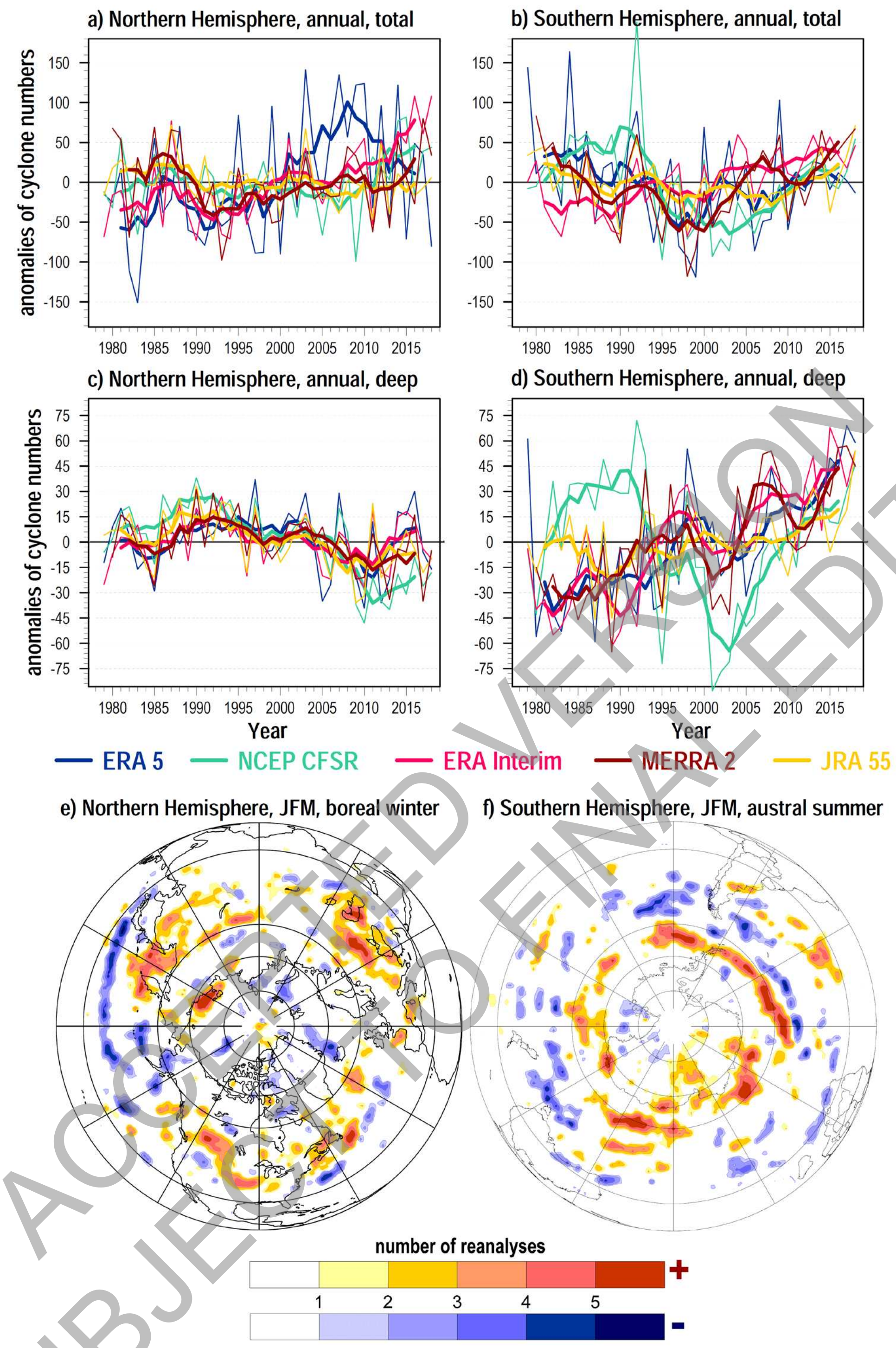


Figure 8.12: Annual anomalies (with respect to the reference period 1979-2018) of the total number of extratropical cyclones (a, c) and of the number of deep cyclones (<980hPa) (b, d) over the Northern (a, b) and the Southern (c, d) Hemispheres in different reanalyses (shown in colors in the legend). Note different vertical scales for panels (a), (b) and (c), (d). Thin lines indicate annual anomalies and bold lines indicate 5-yr running averages. (e), (f) The number of reanalyses (out of five) simultaneously indicating statistically significant (90% level) linear trends of the same sign during 1979–2018 for JFM over the Northern Hemisphere (e) and over the Southern Hemisphere (f). Updated from (Tilinina et al., 2013). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

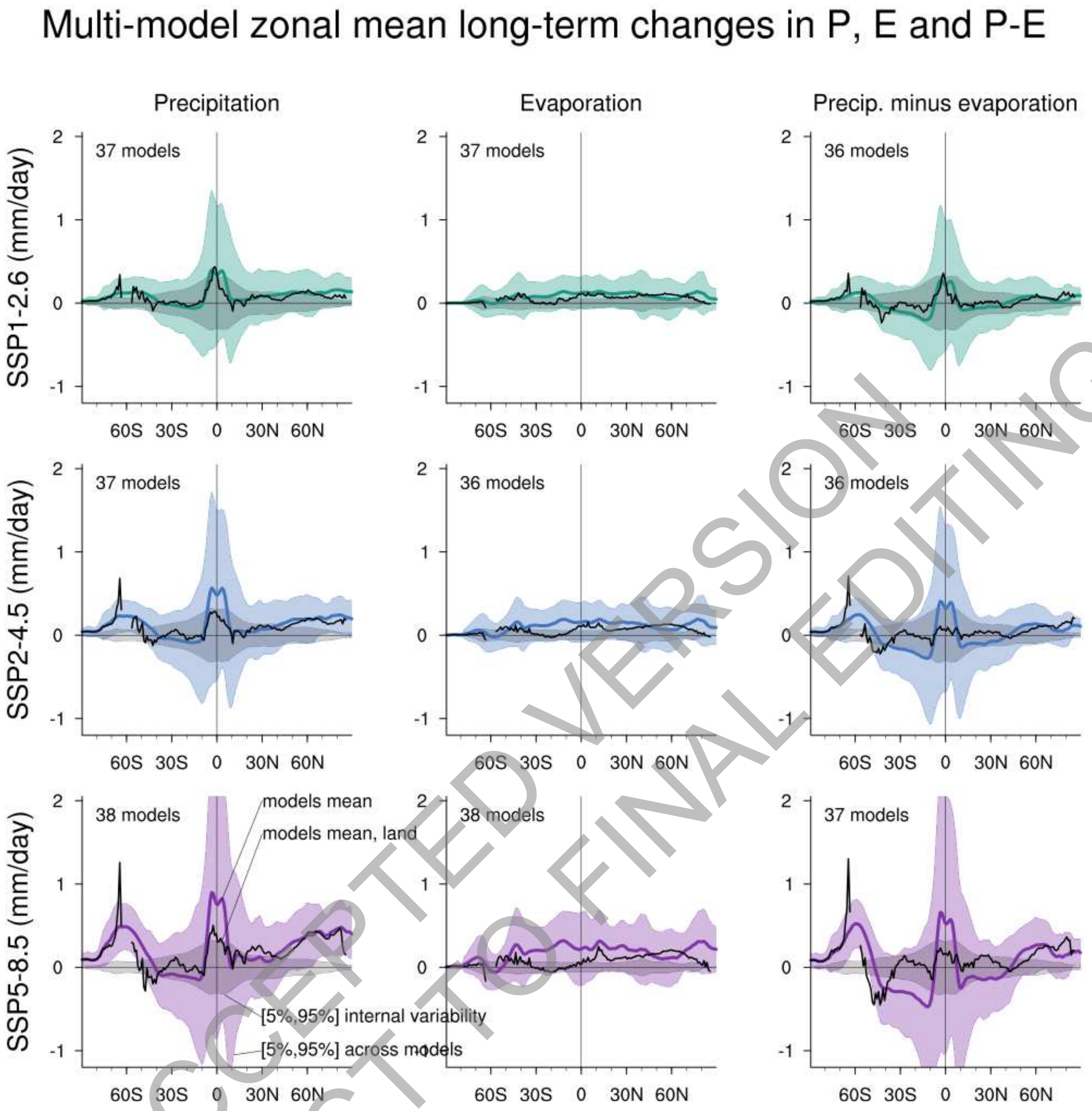


Figure 8.13: Zonal and annual mean projected long-term changes in the atmospheric water budget. Zonal and annual mean projected changes (mm/day) in P (precipitation, left column), E (evaporation, middle column), and P-E (right column) over both land and ocean areas (thick line) and over land only (dashed line) averaged across 36 to 38 CMIP6 models in the SSP1-2.6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario, respectively. Shading denotes confidence intervals estimated from the CMIP6 ensemble under a normal distribution hypothesis. Color shading denotes changes over both land and ocean. Grey shading represents internal variability derived from the pre-industrial control simulations. All changes are estimated for 2081-2100 relative to the 1995-2014 base period. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

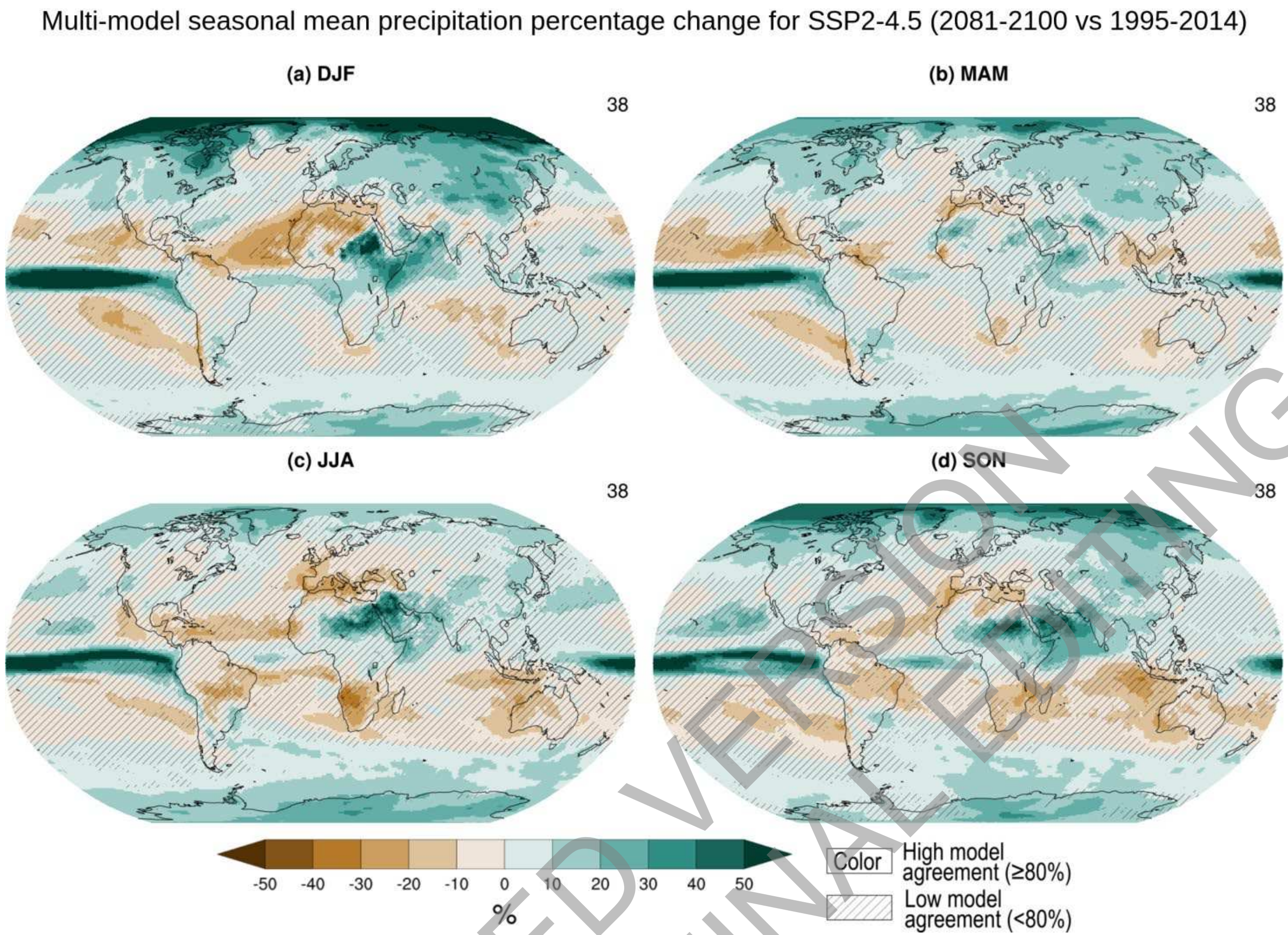
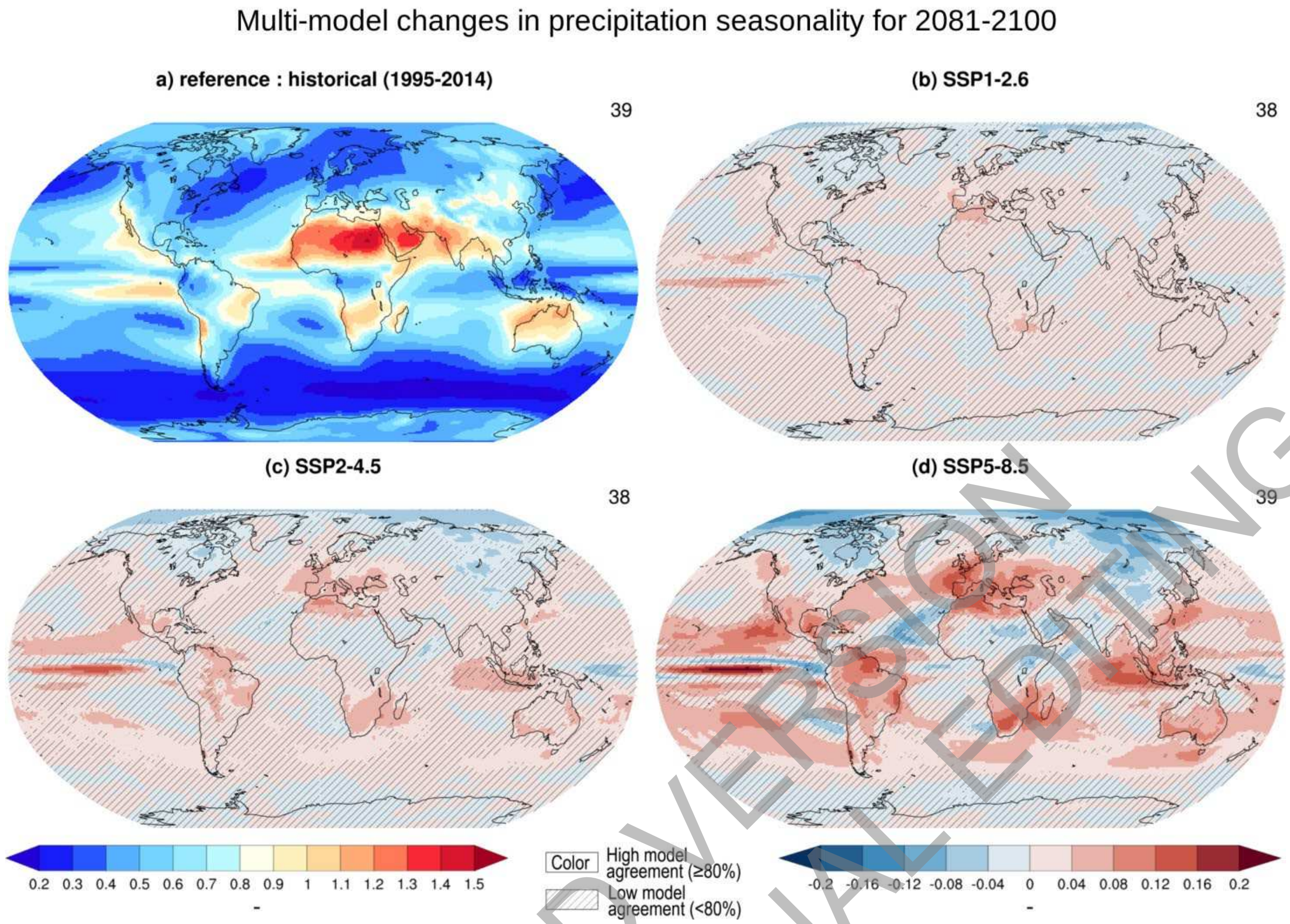


Figure 8.14: Projected long-term relative changes in seasonal mean precipitation. Global maps of projected relative changes (%) in seasonal mean of precipitation averaged across 29 CMIP6 models in the SSP2-4.5 scenario. All changes are estimated for 2081-2100 relative to the 1995-2014 base period. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).



Box 8.2, Figure 1: Projected long-term changes in precipitation seasonality. Global maps of projected changes in precipitation seasonality (simply defined as the sum of the absolute deviations of mean monthly rainfalls from the overall monthly mean, divided by the mean annual rainfall as in Walsh and Lawler, 1981) averaged across 31 to 33 CMIP6 models in the SSP1-2.6 (b), SSP2-4.5 (c) and SSP5-8.5 (d) scenario respectively. The simulated 1995-2014 climatology is shown in panel (a). All changes are estimated in 2081-2100 relative to 1995-2014. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

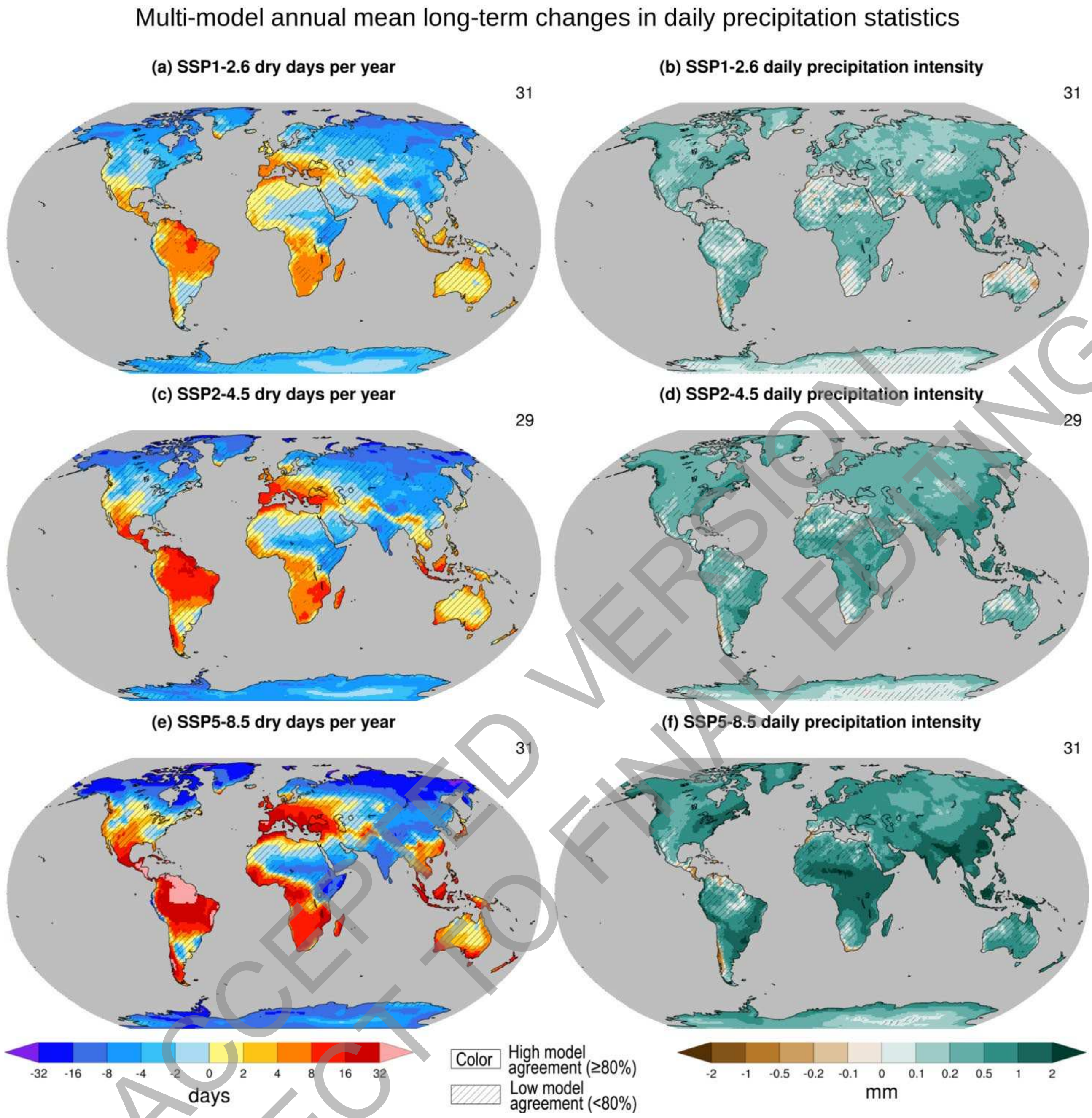


Figure 8.15: Projected long-term relative changes in daily precipitation statistics. Global maps of projected seasonal mean relative changes (%) in the number of dry days (i.e. days with less than 1 mm of rain) and daily precipitation intensity (in mm/day, estimated as the mean daily precipitation amount at wet days - i.e., days with intensity above 1 mm/day) averaged across CMIP6 models in the SSP1-2.6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario respectively. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

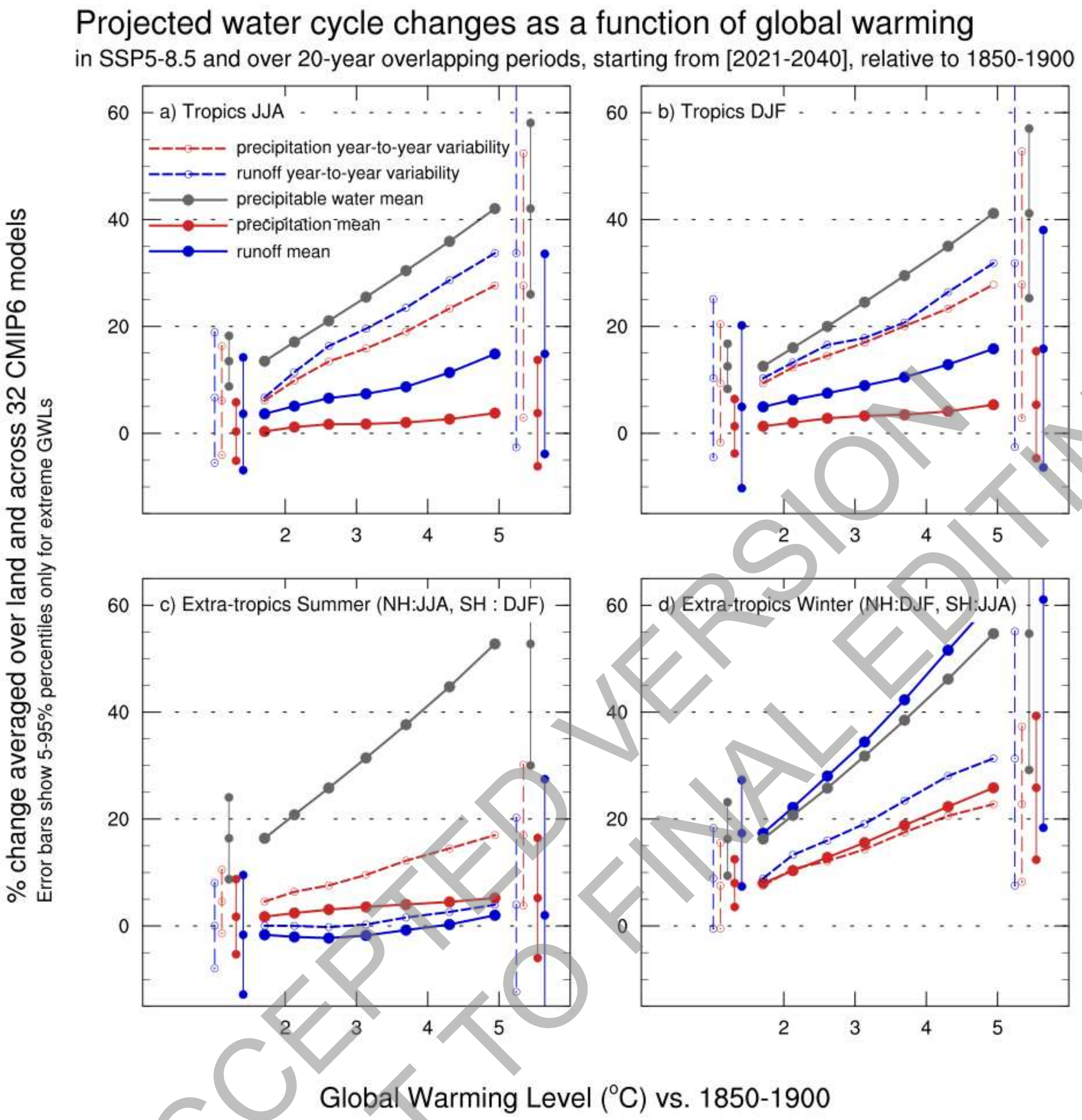


Figure 8.16: Rate of change in mean and variability across increasing global warming levels. Relative change (%) in seasonal mean total precipitable water (green dashed line), precipitation (red dashed lines), runoff (blue dashed lines), as well as in standard deviation of precipitation (red solid lines) and runoff (blue solid lines) averaged over extra-tropical land in (a) summer and (b) winter, and tropical land in (c) JJA and (d) DJF as a function of global-mean surface temperature for the CMIP6 multi-model mean across the SSP5-8.5 scenario. Extra-tropical winter refers to DJF for Northern Hemisphere and JJA for Southern Hemisphere (and the reverse for extra-tropical summer). Each marker indicates a 21-year period centered on consecutive decades between 2015 and 2085 relative to the 1995–2014 base period. Precipitation and runoff variability are estimated by their standard deviation after removing linear trends from each time series. Error bars show the 5-95% confidence interval for the warmest 5°C global warming level. Figure adapted from (Pendergrass et al., 2017) and updated with CMIP6 models. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

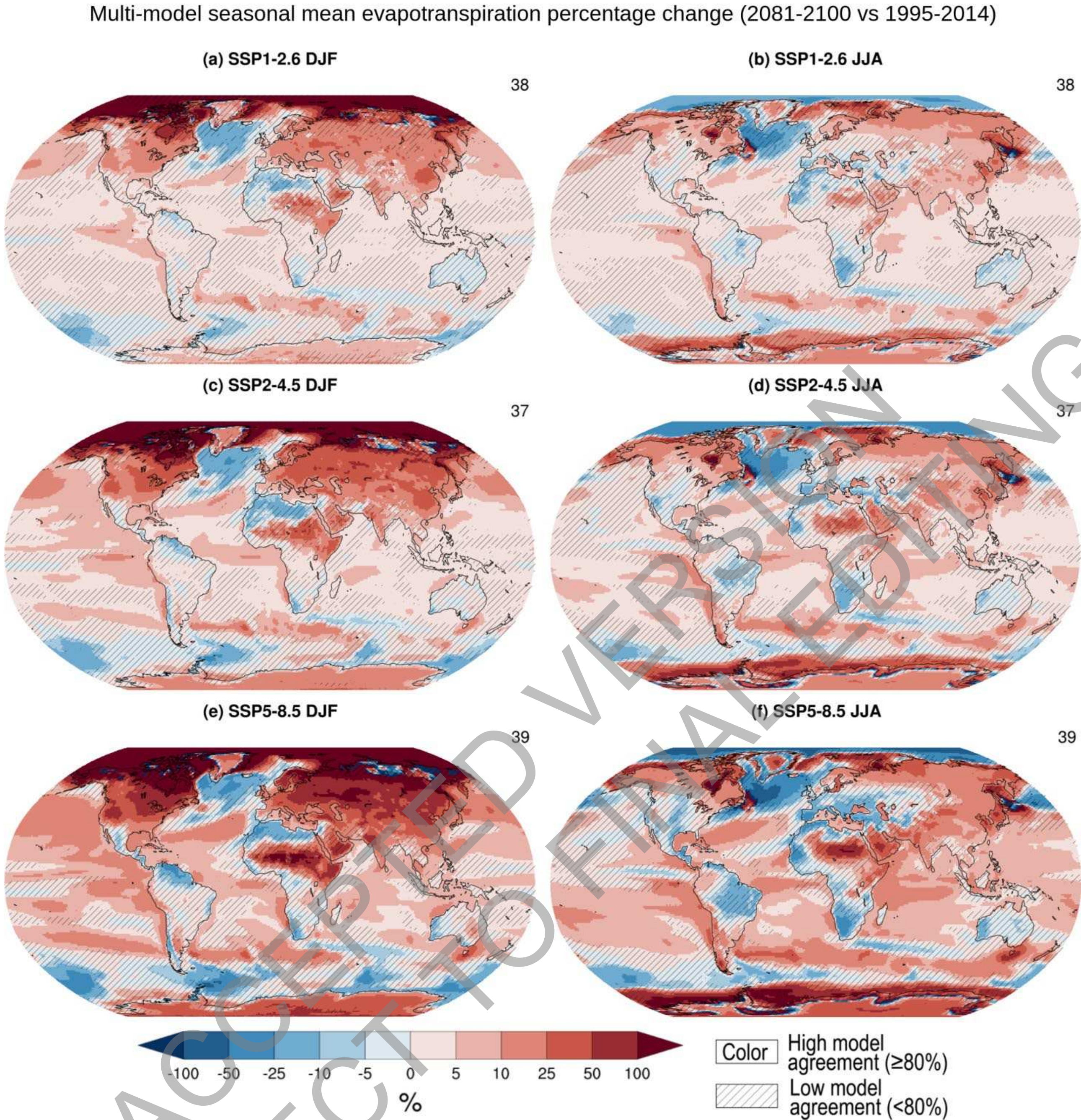


Figure 8.17: Projected long-term relative changes in seasonal mean evapotranspiration. Global maps of projected relative changes (%) in seasonal mean of surface evapotranspiration for DJF (left panels) and JJA (right panels) averaged across 29 or 30 CMIP6 models for SSP1.2-6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario respectively. All changes are estimated in 2081-2100 relative to 1995-2014. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

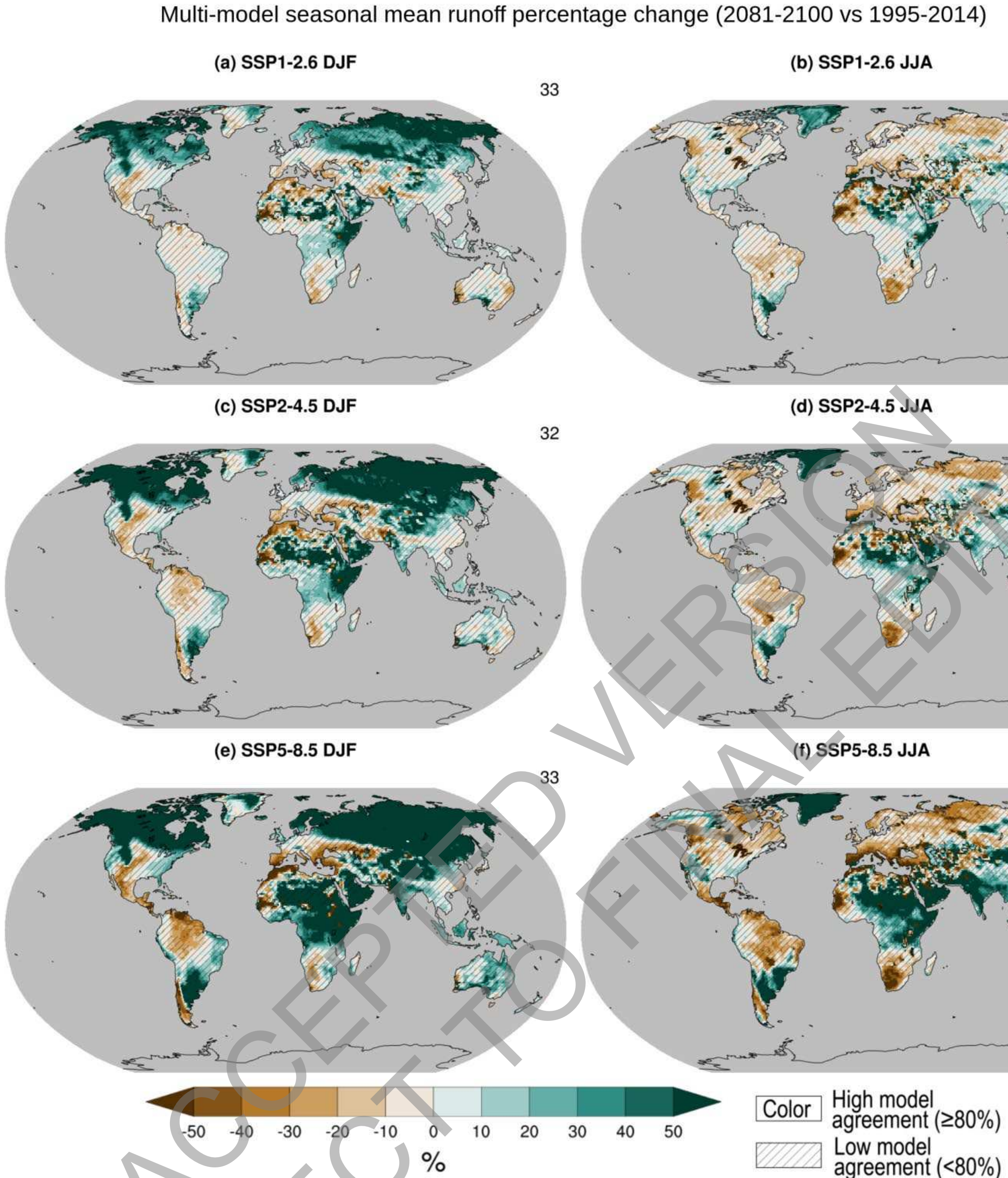


Figure 8.18: Projected long-term relative changes in seasonal mean runoff. Global maps of projected relative change (%) in runoff seasonal mean for DJF (left panels) and JJA (right panels) averaged across CMIP6 models SSP1.2-6 (a,b), SSP2-4.5 (c,d) and SSP5-8.5 (e,f) scenario respectively. All changes are estimated in 2081-2100 relative to 1995-2014. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

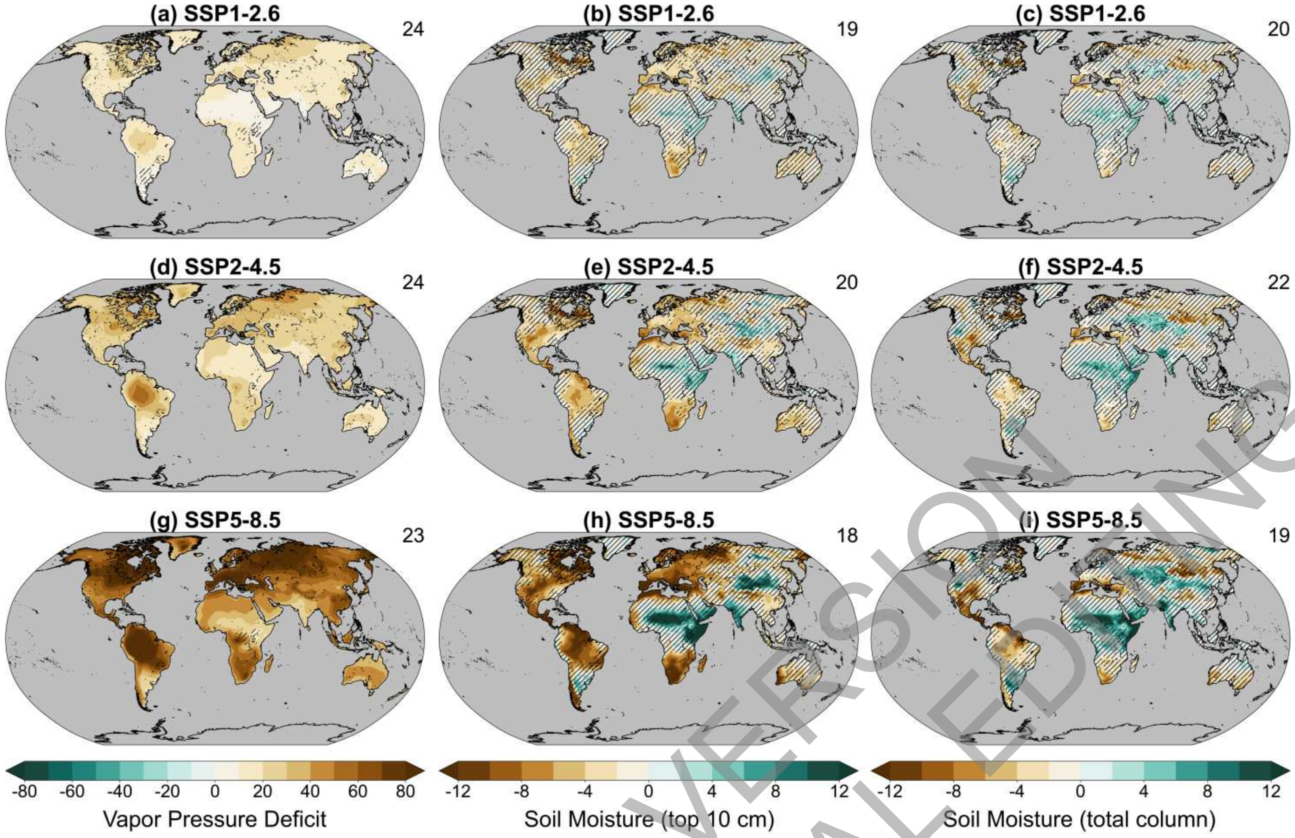


Figure 8.19: Projected long-term relative changes in annual mean soil moisture and vapor pressure deficit. Global maps of projected relative changes (%) in annual mean vapor pressure deficit (left), surface soil moisture (top 10cm, middle) and total column soil moisture (right) from available CMIP6 models for the SSP1.2-6 (a,b,c), SSP2-4.5 (d,e,f) and SSP5-8.5 (g,h,i) scenarios respectively. All changes are estimated for 2081-2100 relative to a 1995-2014 base period. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement ("Robust change"), where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

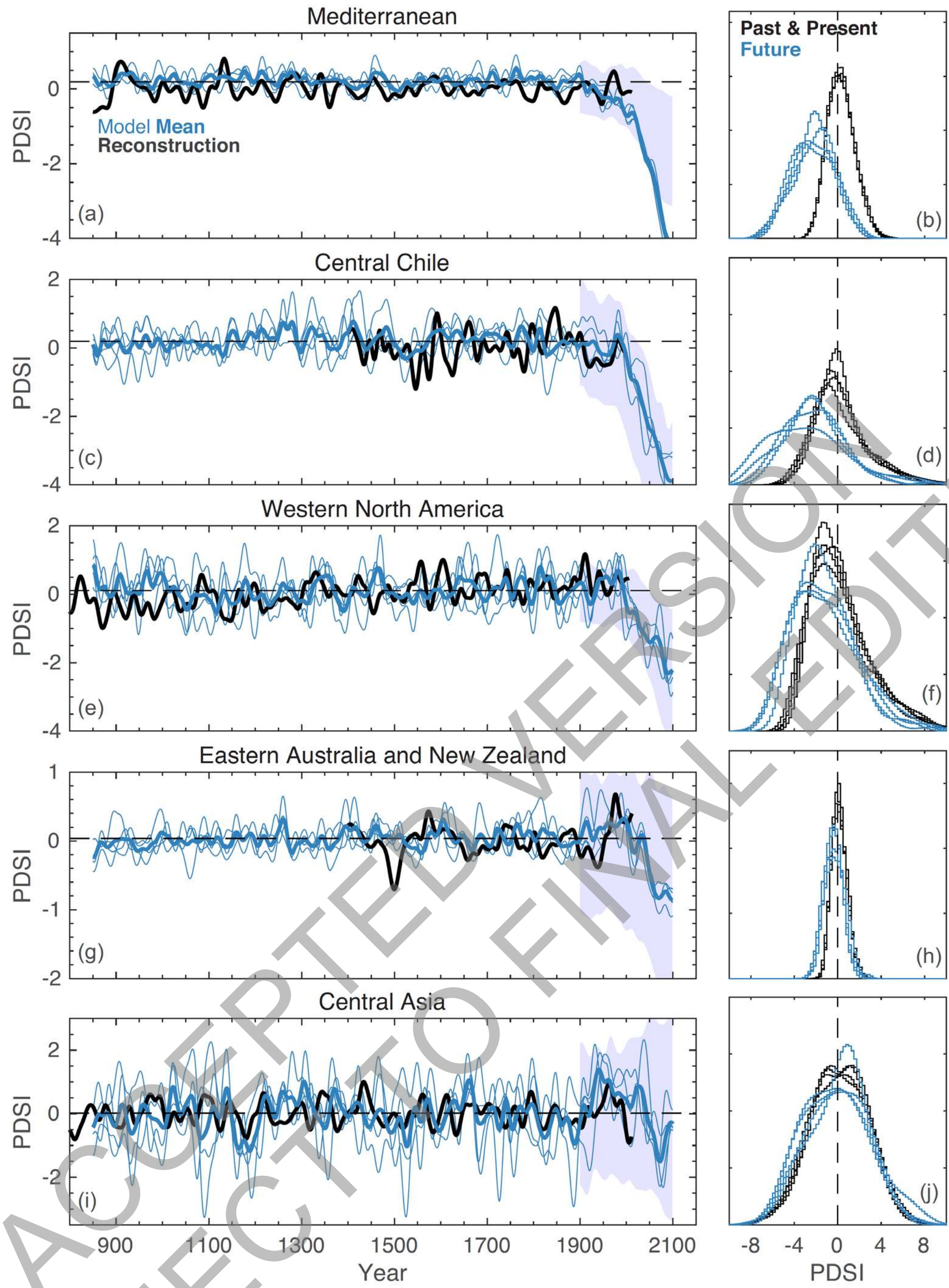


Figure 8.20: Past-to-future drought variability in paleoclimate reconstructions and models for select regions. On the left (a,c,e,g,i), tree-ring reconstructed Palmer Drought Severity Index (PDSI) series (black line) for the Mediterranean (10˚W–45˚E, 30˚–47˚N; Cook et al., 2015, 2016), central Chile (70˚–74˚W, 32˚–37˚S; Morales et al., 2020), western North America (117˚–124˚W, 32˚–38˚N; Cook et al., 2010; Griffin and Anchukaitis, 2014), Eastern Australia and New Zealand (136˚–178˚E, 46˚–11˚S; Palmer et al., 2015), and Central Asia (99˚–107˚E, 47˚–49˚N; Pederson et al., 2014; Hessl et al., 2018) plotted in comparison to the past-to-future fully-forced simulations from four ensemble members (thin blue lines) from the NCAR CESM Last Millennium Ensemble (thick blue line = ensemble mean) (Otto-Bliesner et al., 2016) for the same regions. The shaded area represents the range (10th to 90th percentile) of historical and future (RCP8.5) PDSI (Penman-Monteith) simulations from 15 CMIP5 models and 34 ensemble members for the same regions (1900–2100; Cook et al., 2014). On the right (b,d,f,h,i), the distribution of annual PDSI values from the past and present (850 to 2005 CE) (black) is compared to the future distribution (2006 to 2100 CE) (blue). The distributions show each of the four ensemble members from the CESM LME simulations. The future component of the CESM LME follows the RCP8.5 scenario. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

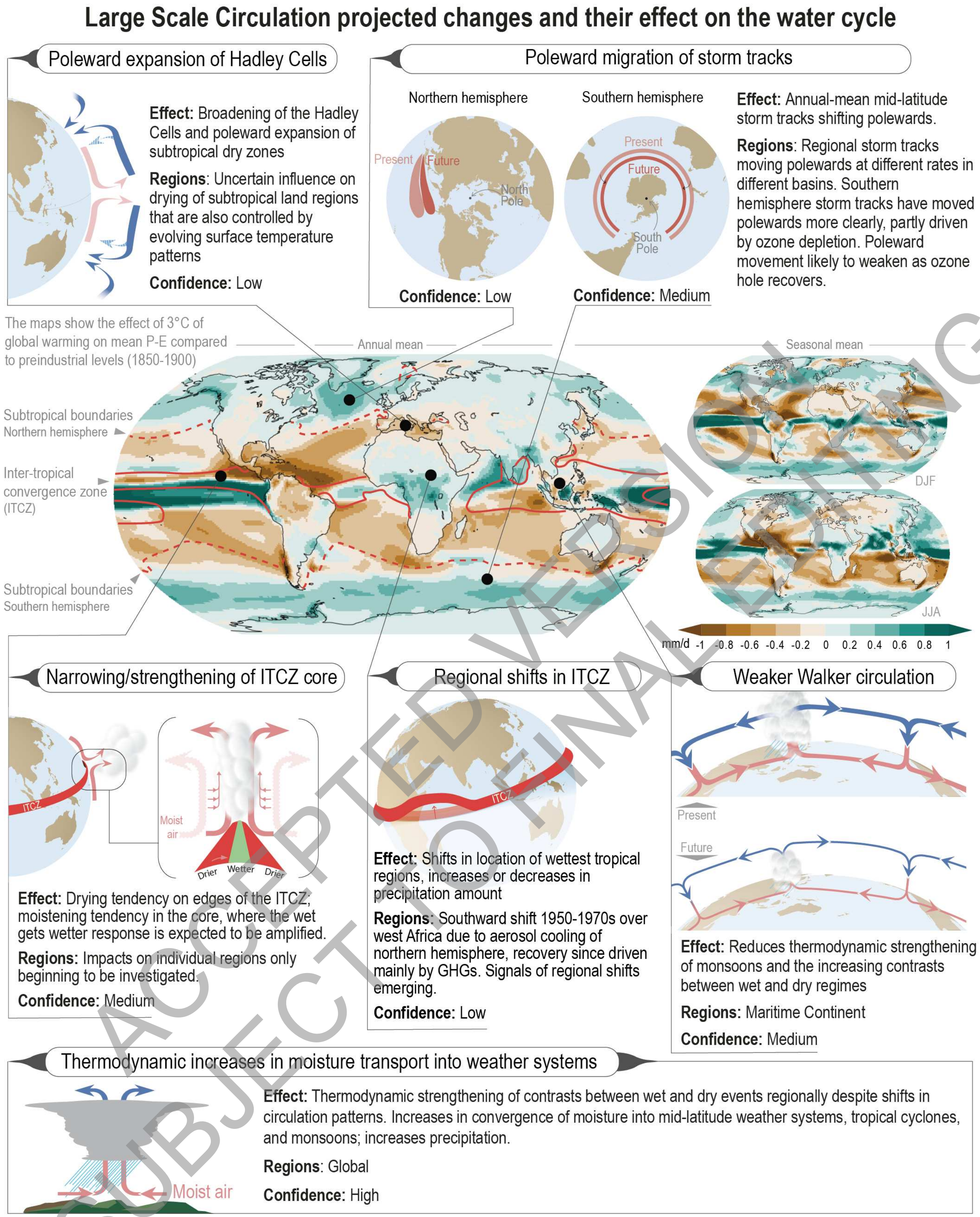


Figure 8.21: Schematic depicting large-scale circulation changes and impacts on the regional water cycle. The central figures show precipitation minus evaporation (P-E) changes at 3°C or global warming relative to a 1850-1900 base period (mean of 23 CMIP6 SSP5-8.5 simulations). Annual mean changes (large map) include contours depicting control climate P-E=0 lines with the solid contour enclosing the tropical rain belt region and dashed lines representing the edges of subtropical regions. Confidence levels assess understanding of how large-scale circulation change affect the regional water.

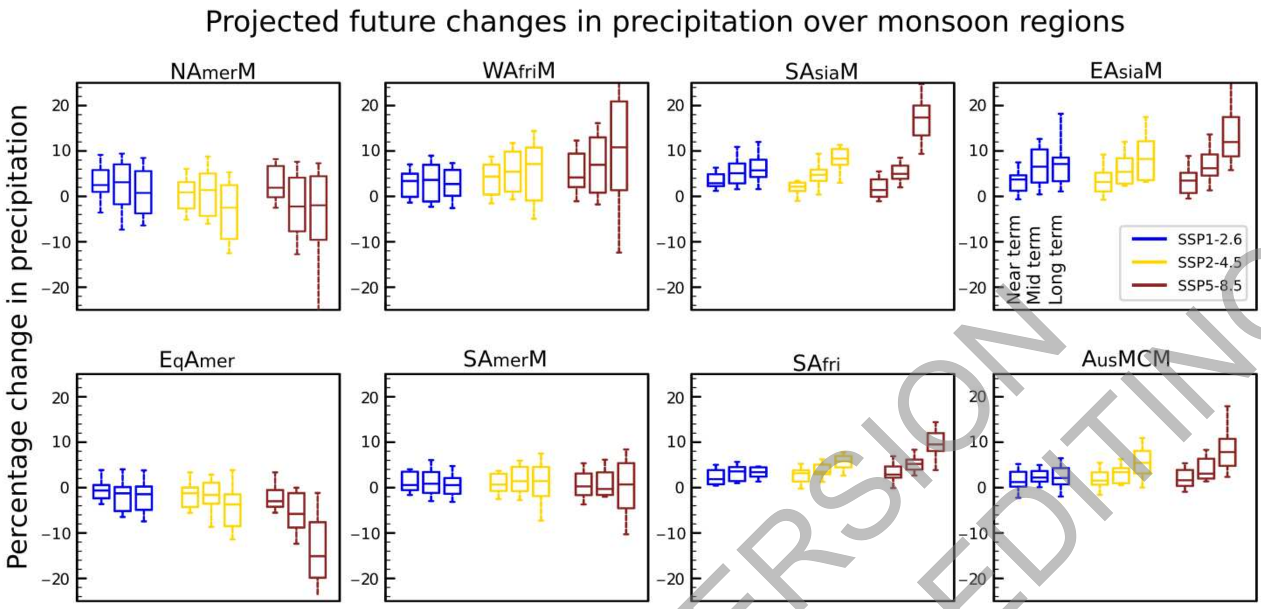


Figure 8.22: Projected regional monsoons precipitation changes. Percentage change in projected seasonal mean precipitation over regional monsoon domains (as defined in Fig 8.11, Section 8.3.2.4 and Annex V) for near-term (2021-2040), mid-term (2041-2060), and long-term (2081-2100) periods based on 24 CMIP6 models and three SSP scenarios (SSP1-2.6, SSP2-4.5 and SSP5-8.5). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

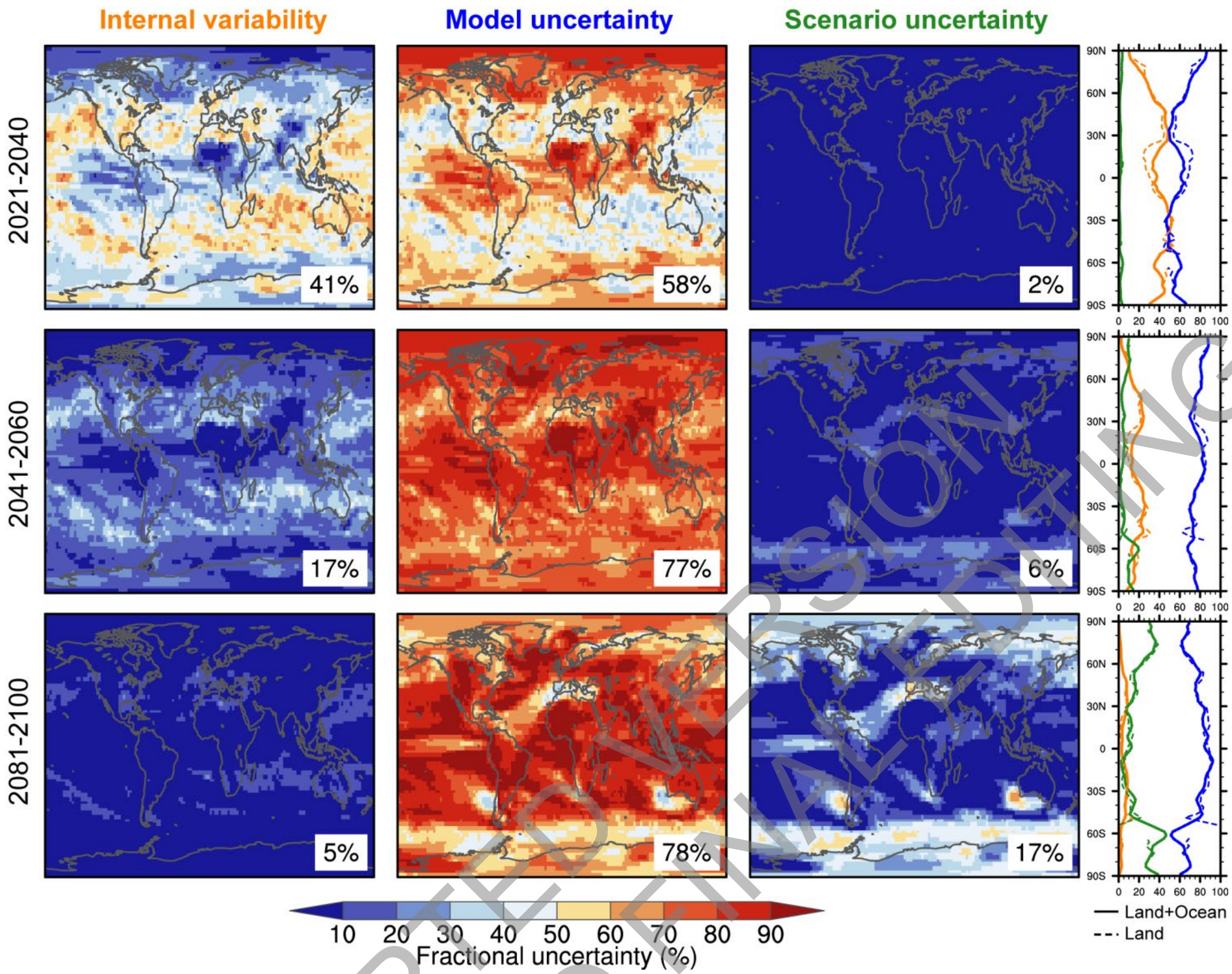


Figure 8.23: Geographical and zonal mean distribution of the percentage of variance explained by the three sources of uncertainty in CMIP6 projections of 20-year mean precipitation changes in 2021–2040 (top), 2041–2060 (middle) and 2081–2100 (bottom) relative to the 1995–2014 base period: internal climate variability (left), model response uncertainty (middle) and scenario uncertainty (right, considering four plausible concentration scenarios: SSP1-2.6, SSP2-4.5, SSP3-7.0 and SSP5-8.5). Percentage numbers give the area-weighted global average value for each map. Right panels show the zonal mean fractions over both land and sea (solid lines) and over land only (dashed line). The figure was adapted from Fig.4a in (Lehner et al., 2020). The relative contributions of internal variability, models and emission scenarios to the total uncertainty depend on both region and time horizon. The scenario uncertainty is relatively low in near and mid-term time horizons while it increases in the long-term mostly over the high-latitudes. The model response uncertainty is the most influential factor across all time horizons. Internal variability also plays a key role in the near-term, especially in the subtropics. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

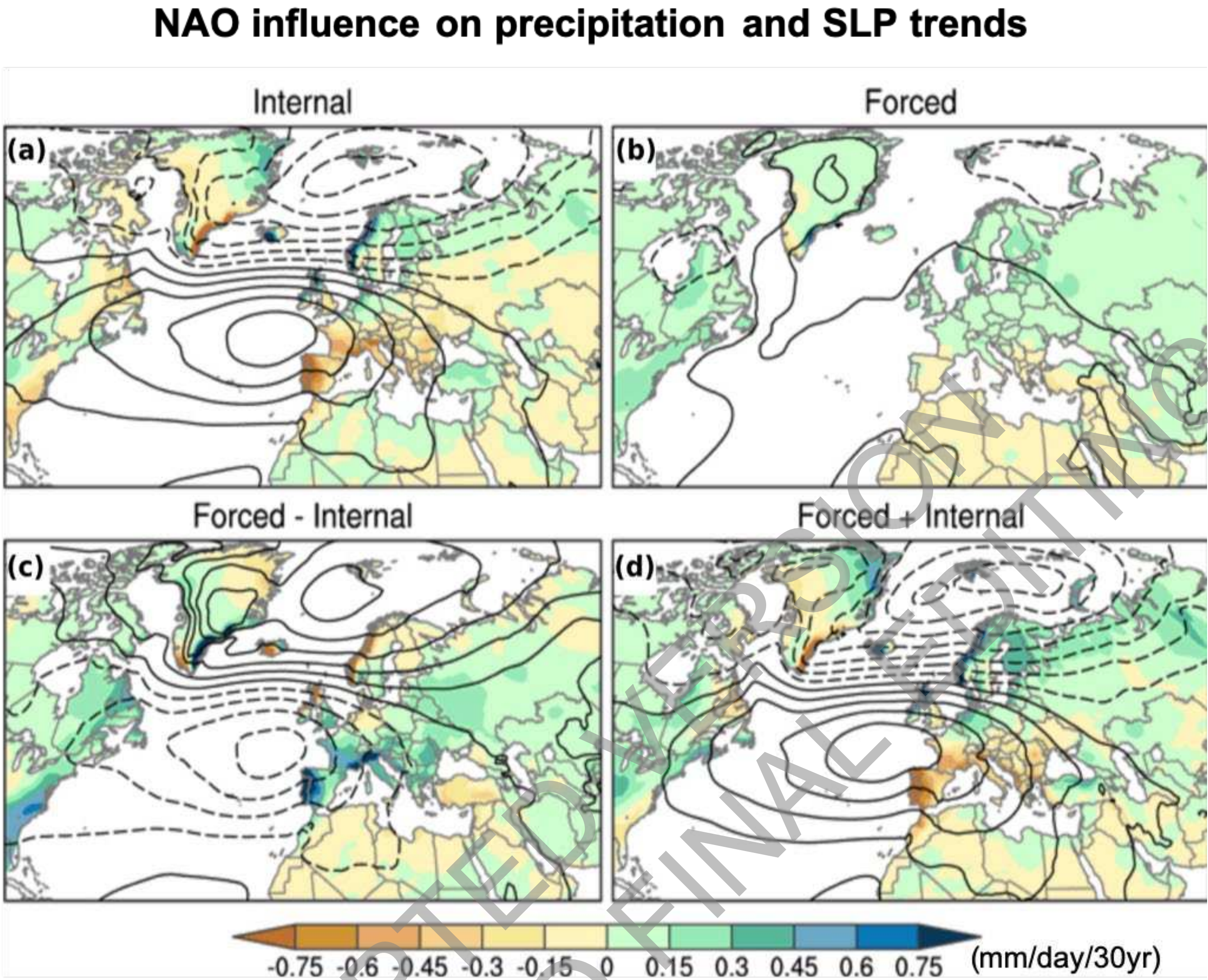


Figure 8.24: Impact of the North Atlantic Oscillation (NAO) on 2016–2045 climate trends. (a) Regressions of winter sea level pressure (SLP) and precipitation trends upon the normalized leading principal component (PC)of winter SLP trends in the CESM1 Large Ensemble, multiplied by two to correspond to a two standard deviation anomaly of the PC (as internal climate variability component); (b) CESM1 ensemble-mean winter SLP and precipitation trends (as forced climate variability component); (c) b – a (forced minus internal climate variability component); (d) b + a (forced plus internal climate variability component). Precipitation in color shading (mm/day per 30 years) and SLP in contours (interval = 1 hPa per 30 years with negative values dashed) (Adapted from Deser et al., 2017). Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

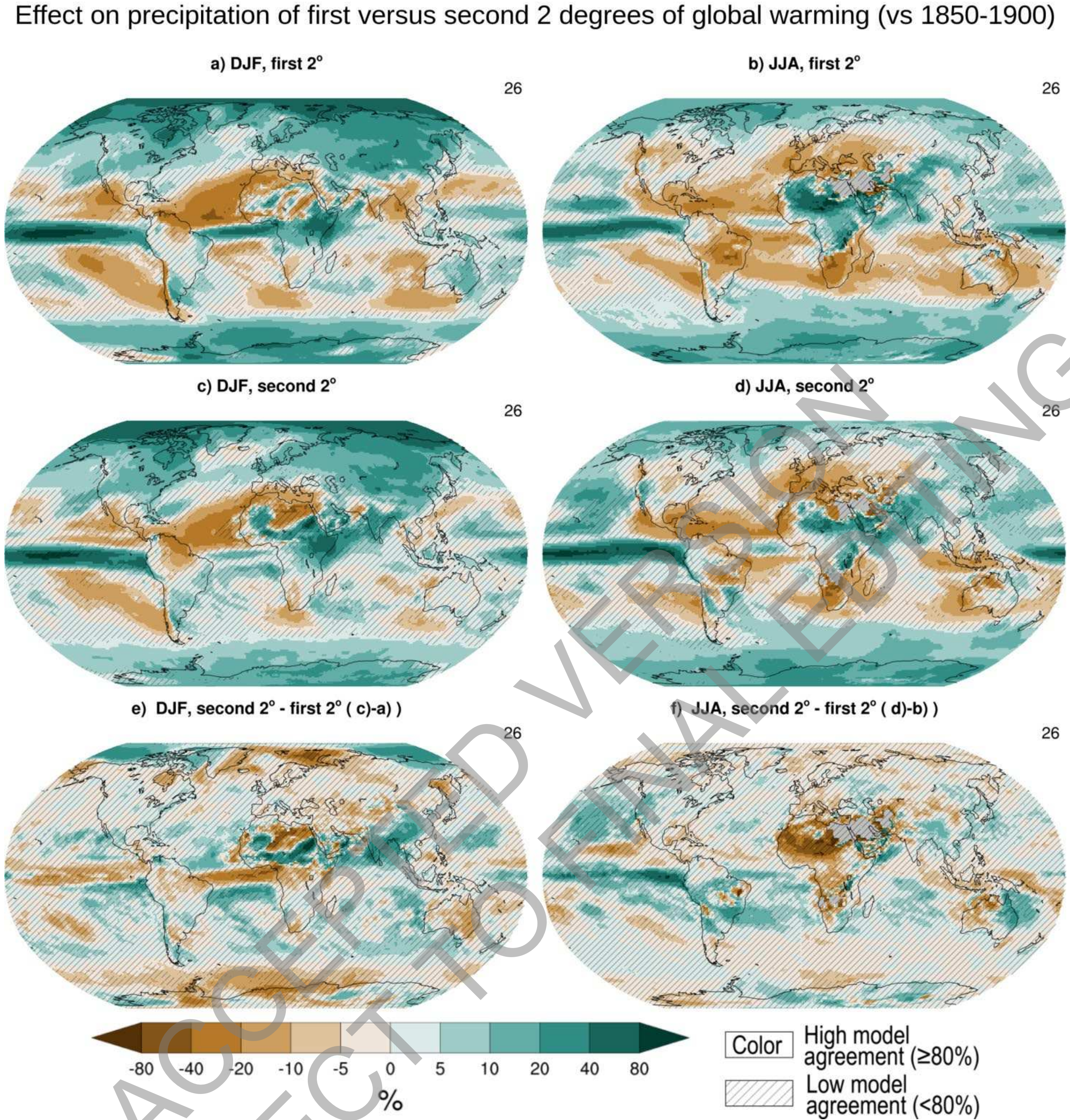


Figure 8.25: Effect of first versus second 2°C of global warming relative to the 1850-1900 base period on seasonal mean precipitation (mm/day). CMIP6 multi-model ensemble mean DJF (left panels) and JJA (right panels) precipitation difference for a,b) SSP5-8.5 at +2°C; c,d) SSP5-8.5 at +4°C minus SSP5-8.5 at +2°C (second 2°C warming); e,f) second minus first 2°C fast warming (c-a and d-b). Only models reaching the +4°C warming levels in SSP5-8.5 are considered. Differences are computed based on 21-yr time windows centered on the first year reaching or exceeding the selected global warming level using a 21-yr running mean global surface atmospheric temperature criterion. Uncertainty is represented using the simple approach: No overlay indicates regions with high model agreement, where ≥80% of models agree on sign of change; diagonal lines indicate regions with low model agreement, where <80% of models agree on sign of change. For more information on the simple approach, please refer to the Cross-Chapter Box Atlas.1. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

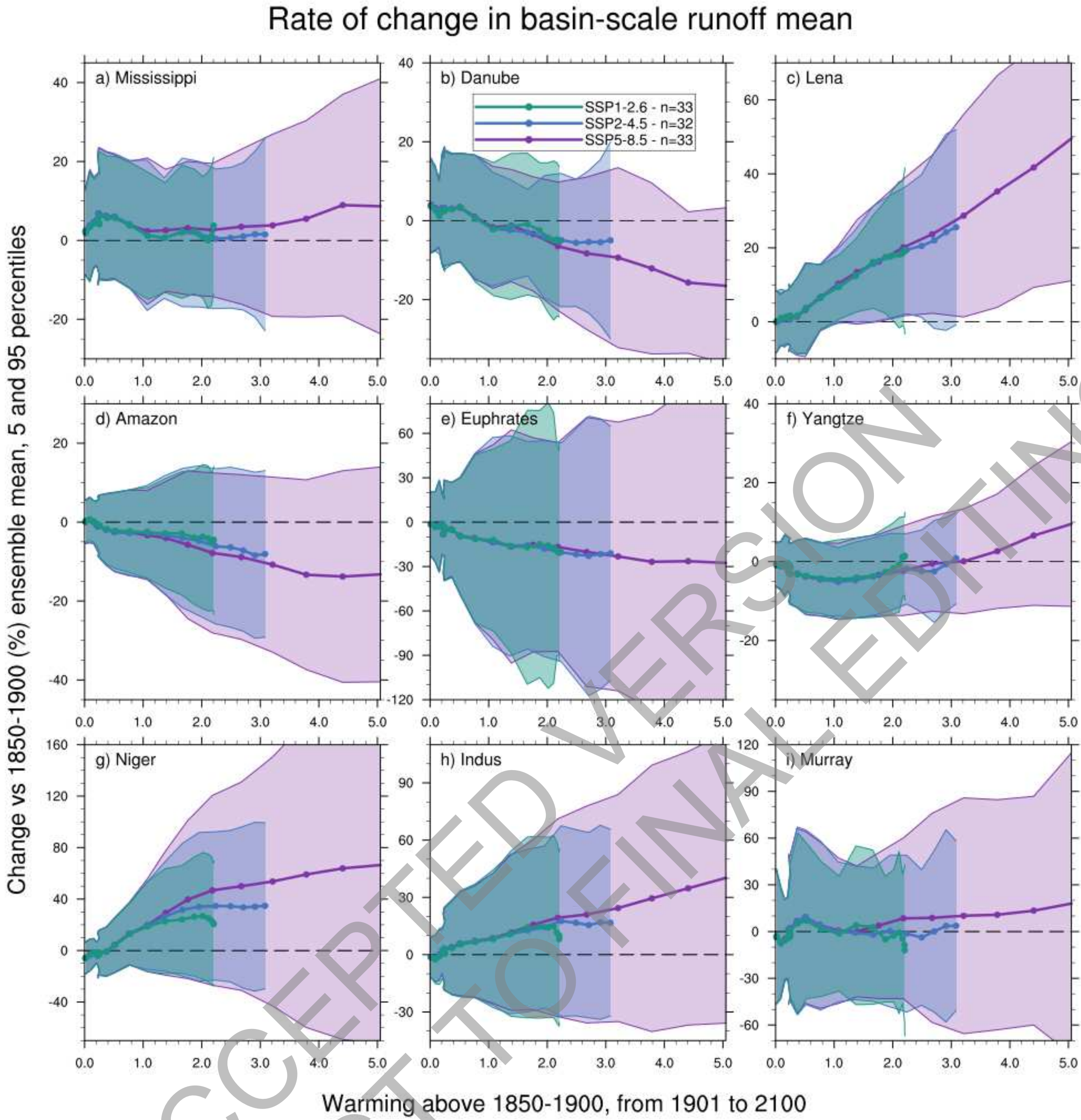


Figure 8.26: Rate of change in basin-scale annual mean runoff with increasing global warming levels. Relative changes (%) in basin-averaged annual mean runoff estimated as multi-model ensemble median from a variable subset of CMIP6 models for each SSP over six major river basins: a) Mississipi, b) Danube, c) Lena, d) Amazon, e) Euphrates, f) Yangtze, g) Niger, h) Indus, i) Murray. The basin averages have been estimated after a first-order conservative remapping of the model outputs on the 0.5° by 0.5° river network of (Decharme et al., 2019). The shaded area indicates the 5-95% confidence interval of the ensemble values across all SSPs. Note that the y-axis range differs across basins and is particularly large for Niger and Murray (panels g and i). The number of models considered is specified for each scenario in the legend located inside panel b. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).

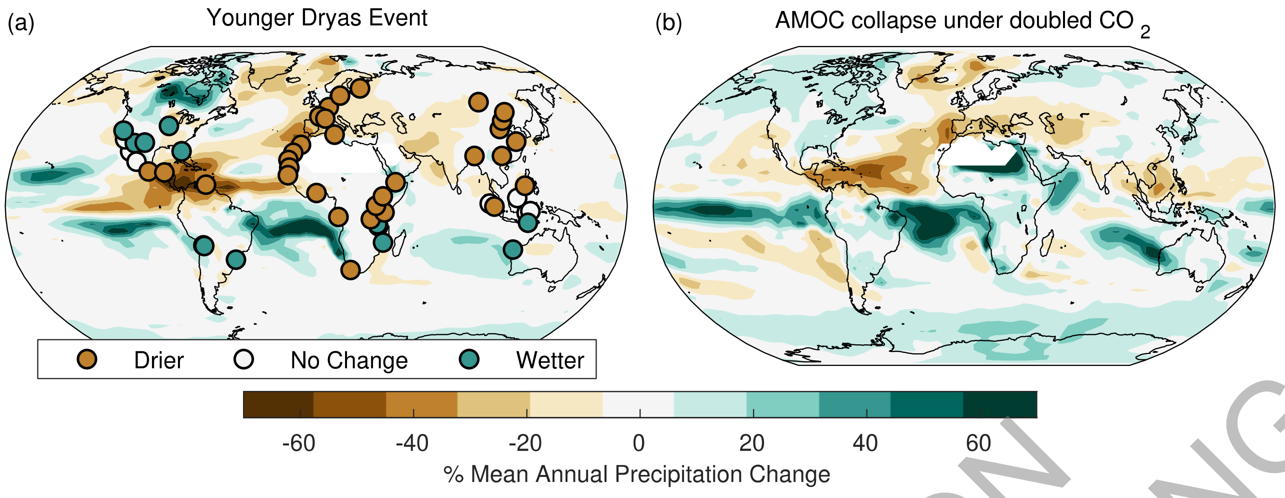
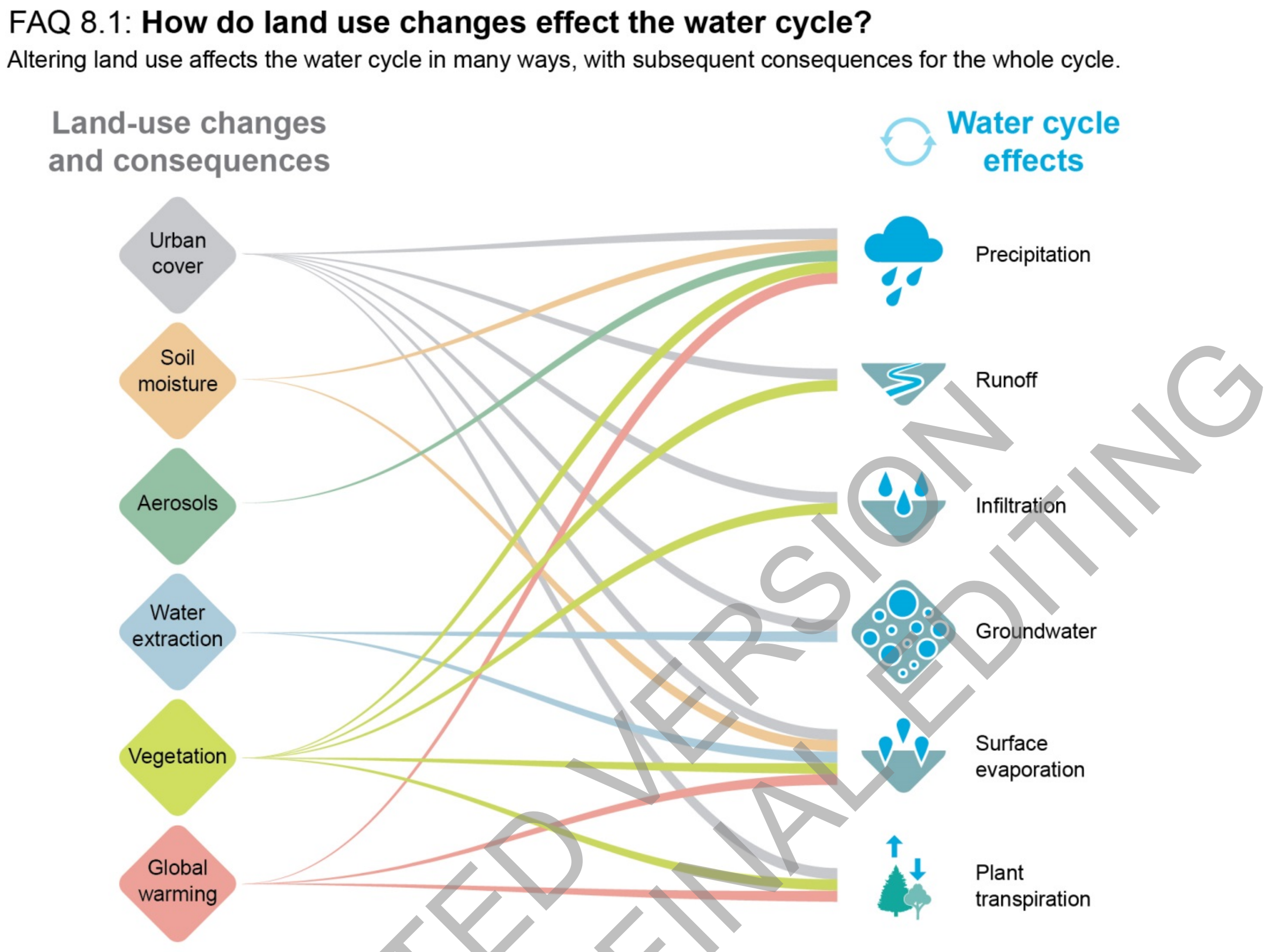
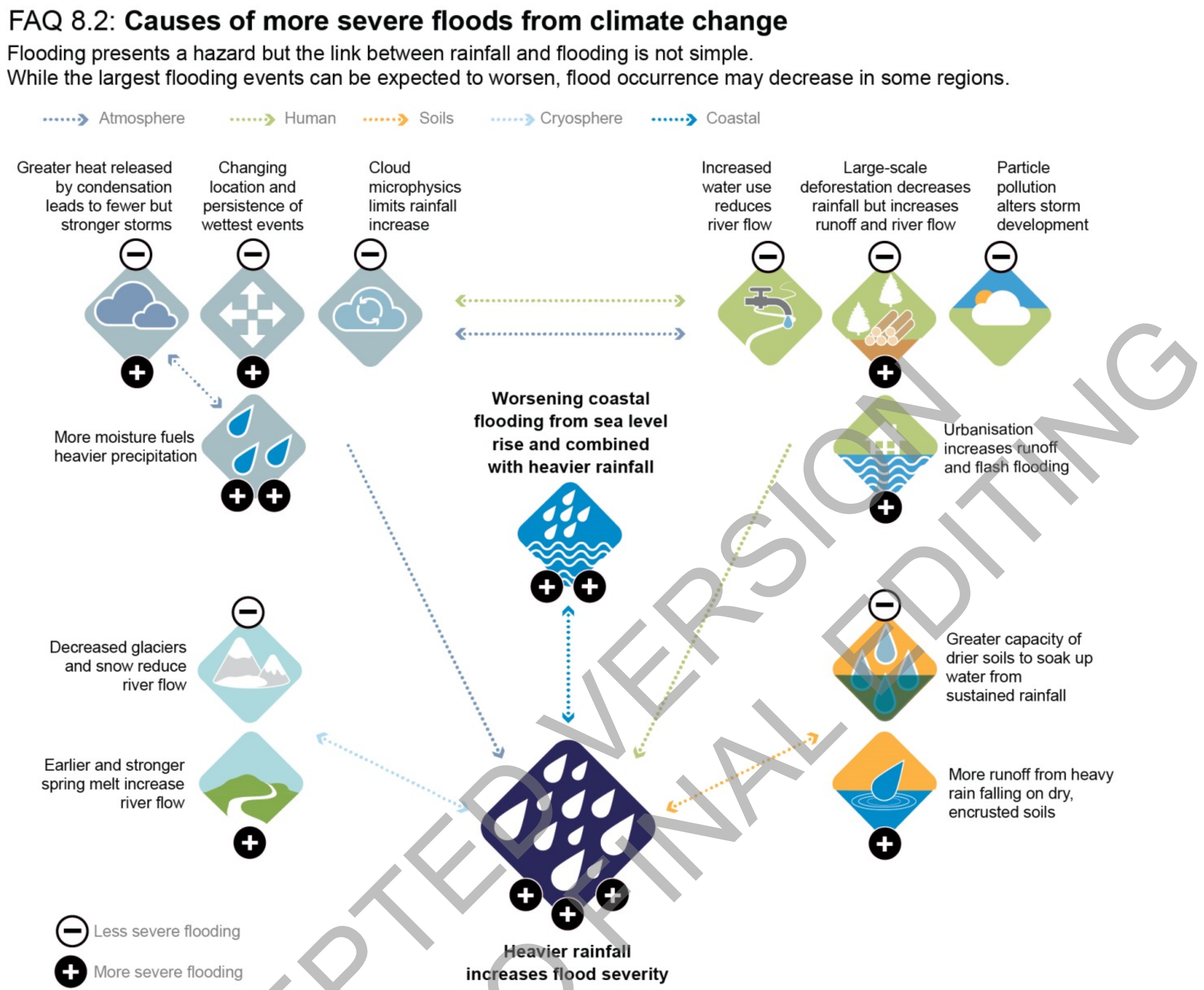


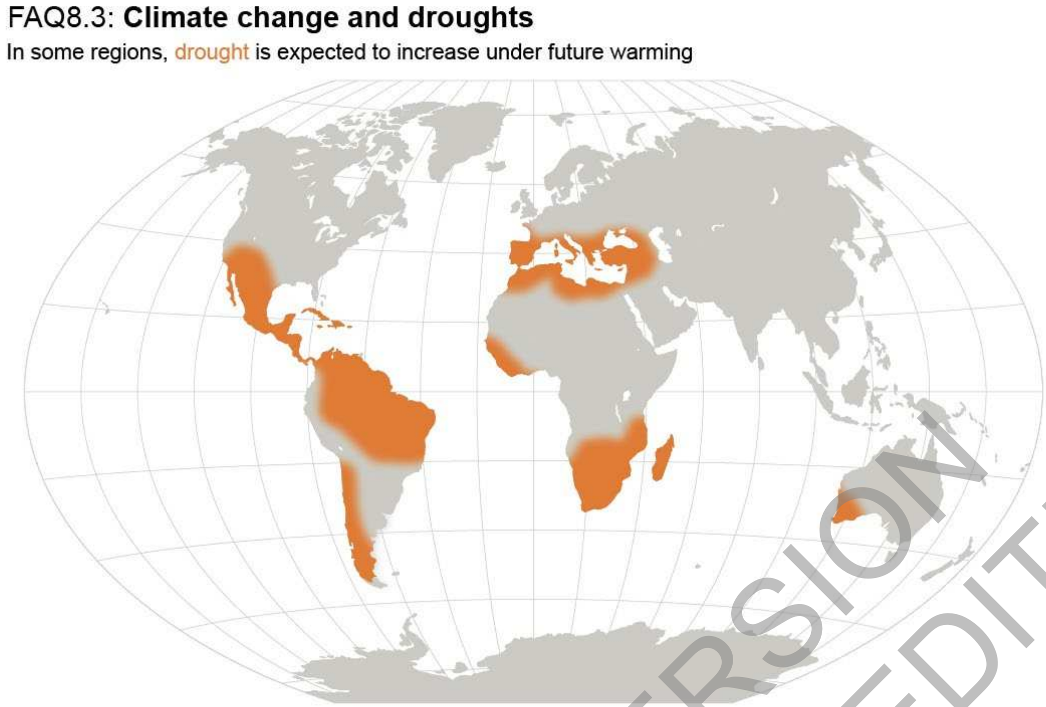
Figure 8.27: (a) Model simulation of precipitation response to the Younger Dryas event relative to the preceding warm Bølling-Allerød period (base colors, calculated as the difference between 12,600–11,700 yr BP and 14,500–12,900 BP from the TraCE paleoclimate simulation of Liu et al., (2009)), with paleoclimate proxy evidence superimposed on top (dots). (b) Model simulation of precipitation response to an abrupt collapse in AMOC under a doubling of 1990 CO_2 levels (after Liu et al., (2017)). Regions with rainfall rates below 20 mm/year are masked. Further details on data sources and processing are available in the chapter data table (Table 8.SM.1).



FAQ8.1, Figure 1: Land-use changes and their consequences on the water cycle. As all the components or the water cycle are tightly connected, changes in one aspect of the cycle affects almost all the cycle.



FAQ 8.2, Figure 1: Schematic illustrating factors important in determining changes in heavy precipitation and flooding.



FAQ 8.3, Figure 1: Schematic map highlighting in brown the regions where droughts are expected to become worse as a result of climate change. This pattern is similar regardless of the emissions scenario; however, the magnitude of change increases under higher emissions.